Bulletin
of
The Geological Society of America

VOLUME 55 JULY 1944 NUMBER 7

CONTENTS

	PAGES
Problems of terrace correlation. By Douglas Johnson	793–818
Geology of the north-central Wasatch Mountains, Utah. By A. J. Eardley	819–894
Mt. Waldo batholith and associated igneous rocks, Waldo County, Maine. By Joseph M. Trefethen	895–904
Petrology of the Pawtuckaway Mountains, New Hampshire. By Chalmer J. Roy and Jacob Freedman	905–920

Subscription, $10 per year.

Publication Office: Mt. Royal and Guilford Aves., Baltimore 2, Md.

Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117th Street, New York 27, N. Y.

NOTICE.—In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter in the Post-Office at Baltimore, Md., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103, Act of October 3, 1917, authorized on July 8, 1918.

PAPERS IN PRESS FOR FORTHCOMING ISSUES

STRATIGRAPHY AND STRUCTURE OF WEST-CENTRAL VERMONT. By Wallace M. Cady

GEOLOGY OF THE SAMOAN ISLANDS. By Harold T. Stearns

PETROLOGY OF THE SAMOAN ISLANDS. By G. A. Macdonald

CORRELATION CHARTS PREPARED BY THE COMMITTEE ON STRATIGRAPHY OF THE NATIONAL RESEARCH COUNCIL, CARL O. DUNBAR, *Chairman*

CORRELATION OF THE CRETACEOUS FORMATIONS OF THE GREATER ANTILLES, CENTRAL AMERICA, AND MEXICO. By Ralph W. Imlay

CORRELATION OF THE CAMBRIAN FORMATIONS OF NORTH AMERICA. By B. F. Howell, *Chairman*

CONTACT METAMORPHISM AT RYE PATCH, NEVADA. By C. J. Vitaliano

GEOLOGY OF THE BARCO CONCESSION, COLOMBIA, SOUTH AMERICA. By F. B. Notestein, C. W. Hubman, and J. W. Bowler

MORRISON FORMATION AND RELATED DEPOSITS IN AND ADJACENT TO THE COLORADO PLATEAU. By William Lee Stokes

RECONSIDERATION OF THE MORRISON FORMATION IN THE TYPE AREA, JEFFERSON COUNTY, COLORADO. By W. A. Waldschmidt and L. W. LeRoy

CORRELATION OF THE DRIFTS OF THE SALAMANCA RE-ENTRANT, NEW YORK. By Paul MacClintock and Earl T. Apfel

PORPHYRY OF THE PORCUPINE DISTRICT, ONTARIO. By J. E. L. Evans

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 55, PP. 793-818, 1 PL., 1 FIG. JULY 1944

PROBLEMS OF TERRACE CORRELATION

BY DOUGLAS JOHNSON*

CONTENTS

	Page
Introduction	793
Coastal-terrace correlation	795
General discussion	795
Possibilities of misinterpretation	795
Criteria of origin	797
Determining altitudes of terraces	799
Difficulties	799
Use of the barometer	799
Significant parts of river terraces	800
Significant parts of coastal terraces	801
Correlating terrace altitudes	802
Logical basis of correlation by altitude	802
Necessity of precise altitude figures	804
Correlation of adjusted values	806
Limitations of the method of correlation by altitude	809
Essential requirements of correlation by altitude	814
Drawing conclusions from terrace correlation	815
Summary	818

ILLUSTRATIONS

Figure	Page
1. Terraces of Coast *X* and Mediterranean Coast plotted to scale	813

Plate	Facing page
1. Shoshone River terraces, Wyoming	800

INTRODUCTION

In the course of geomorphic studies the writer has had occasion to discuss in the field with many investigators certain problems of river-terrace and coastal-terrace correlation. Among experiences most helpful in formulating his ideas may be mentioned excursions to the Rhone valley with Deperet; to the Central Plateau of France with Baulig; to the Seine and neighboring valleys of the Paris Basin with Chaput, Briquet, De Martonne, and others; to the shores of western and southeastern Australia with Clarke, Jutson, Sir Edgeworth David, Leo Cotton, Richards, and other colleagues from the universities at Perth, Adelaide, Melbourne, Sydney, and Brisbane; to the terraced valleys and coasts of New Zealand with C. A. Cotton, Bartrum, Henderson, Speight, Jobberns, Sylvester, and others; to the shores of southeastern Japan with Uchida and other colleagues of the University of Tokyo; around the shores of Oahu with Palmer and other colleagues of the University of Hawaii. These excursions in the company of coworkers have been supplemented by personal visits to a number of localities on the west coast of Scotland, western

* Deceased, February 24, 1944. Professor P. F. Kerr and Mr. Arthur Strahler of Columbia University kindly checked and assembled the virtually completed paper for publication.

Norway, the Baltic shores of Germany and Sweden, and parts of Algeria and South Africa, in the course of which special attention was given to evidences of changes of level recorded in valley and coastal forms.

In this country the writer has profited from more or less extended field discussions of terrace problems with many who have prosecuted special terrace studies. Among others may be mentioned Barrell for the supposed marine terraces of New England and the Piedmont; Florence Bascom, E. B. Knopf and Anna Jonas for terraces of the Piedmont and adjacent regions; Russell and Fisk for terraces of central and southern Louisiana; Barton and Price for coastal terraces of Texas; Leverett, Cooke, and others for coastal terraces of the Atlantic and Gulf slopes; Flint for fluvioglacial terraces of the Connecticut and adjacent valleys; Fuller for many parts of the New England coast. It has also been the writer's fortune to direct intensive terrace studies prosecuted in various parts of this country by a number of his advanced students. As President of the Commission for the Study of Pleistocene and Pliocene Terraces (International Geographical Union) he gave much attention to the work on terrace studies being carried forward in different parts of the world. With the aid of funds provided by Columbia University he made a trip around the world to study, among other things, the origin of low coastal platforms widely (and apparently incorrectly) interpreted as proof of a recent world-wide eustatic drop of sea level. Aided by funds made available through the American Geographical Society, and further assisted by Federal, State, and individual co-operation, he traversed the Atlantic and Gulf coasts for the special purpose of determining the causes of widely divergent terrace correlations for those areas.

The preceding paragraphs do not exhaust the list of those to whom the writer is indebted for valuable aid in terrace studies, but they suffice to show that the conclusions stated in the following pages result from many years of thought and discussion of terrace-correlation problems, thought and discussion to which a great number of investigators have contributed. Some aspects of marine-terrace correlation have been presented in earlier papers.[1] The object of the present paper is to treat, more fully than has before been attempted, some of the fundamental principles upon which successful correlation of both river terraces and coastal terraces must be based. Grateful acknowledgment is made to the many colleagues and students, whether or not specifically named in these pages, whose active discussions, sometimes *pro* and sometimes *con*, have helped the writer in formulating what is here set forth.

In these pages no attempt is made to establish the correctness or the incorrectness of any particular terrace correlation. Actual instances of correlation are cited only when this will the better illustrate or clarify some principle of correlation. To keep discussion impersonal, and to concentrate attention on principles rather than on persons and places, citation of actual correlations will remain anonymous.

Attention is first directed to certain aspects of coastal-terrace correlation, and throughout the text this type of terrace correlation is given the major attention it

[1] Johnson, Douglas (1932) *Principles of marine level correlation*, Geog. Rev., vol. 22, p. 294–298.
——— (1933) *The correlation of ancient marine levels*, Cong. Inter. Géog., Paris, 1931, C. R., Tome II, p. 42–54, Paris.
——— (1933) *Supposed two-meter eustatic bench of the Pacific shores*, Cong. Inter. Géog., Paris, 1931, C. R., Tome II, p. 158–163, Paris.
——— (1936) *Terrace studies in the U. S.*, Cong. Inter. Géog., Varsovie, 1934, C. R., Tome II Travaux de la Section II, 19 p., Warsaw.

seems to merit. But problems of river-terrace correlation, discussed by many authors as throwing light on coastal-terrace correlation, will be treated to the extent deemed appropriate.

COASTAL-TERRACE CORRELATION

GENERAL DISCUSSION

The correlation of coastal terraces appears one of the simplest and most fascinating of geomorphic problems. Four apparently easy steps lead directly to conclusions of great importance. One (1) observes a series of terraces bordering the sea at various localities along one or several coasts, (2) determines the height above sea level of the different terraces at a sufficient number of points, (3) correlates the terraces on the basis of the altitude figures, and (4) from such correlation draws conclusions in favor of local differential land movements, widespread epeirogenic land movements, or world-wide variations in sea level, as the case may be. For so simple a study no special training or experience seems necessary, no unusual precautions essential to success. With reasonable care and sufficient labor any competent geologist or geographer should secure results of much value to others engaged in similar studies.

In view of this situation it is not surprising that great numbers of geologists and geographers in many lands have attempted correlation of coastal terraces. The surprising thing, to those who believe the problem simple, is the remarkable diversity of conclusions reached. There is an enormous literature on the subject of terrace correlation, but the picture obtained by those who critically review this literature is one of confusion and uncertainty. Evidently the problem is more difficult than it appears.

The correlation of coastal terraces is, in fact, one of the most difficult and delicate of geomorphic operations. Every one of the four steps outlined is beset with serious difficulties. Extraordinary precautions must be taken to avoid numerous possibilities of error. To overcome difficulties and guard against errors the worker in this field must make special preparation for his study. He must, of course, equip himself with a thorough knowledge of normal shoreline morphology. But he must do more than this. He must acquaint himself with the origin and evolution of landforms produced by agents other than waves and currents, particularly with forms due to stream action.

The necessity of such preparation should be obvious. Streams, as well as waves and currents, operate in the coastal zone where terraces are to be studied. To embark on such study without the ability to discriminate fluvial from marine forms is clearly hazardous. The discrimination in question is not always easy. Indeed, as we shall show, it is often extremely difficult, even for the expert morphologist. And if the morphologist finds it difficult, how much more the geologist and geographer who have specialized in other fields. Many studies of coastal terraces have been invalidated from the start because the investigator did not think it necessary to determine whether the terraces observed by him were really of marine origin.

POSSIBILITIES OF MISINTERPRETATION

To observe a series of terraces bordering the sea, and to proceed to correlate them on the assumption that they are of marine origin, is manifestly a hazardous pro-

cedure. The origin of the terraces is a matter for specific determination. Terraces may be produced in a great variety of ways, and many types have no significance in fixing former sea levels. To illustrate this point, attention is here directed to a single type of terrace, frequently developed both inland and on coastal borders, which, when observed on coasts, is in danger of being erroneously ascribed to wave action during a higher stand of sea level.

The writings of Gilbert, McGee, Lawson, Paige, Bryan, Davis, and others have demonstrated that in semiarid and arid regions mountain ranges are commonly bordered by rock planes of remarkable perfection, sometimes bare and sometimes veneered with stream-deposited alluvium. The present writer, expanding an early idea of Gilbert's, has shown that these rock planes or "pediments" will result whenever the load of a stream is sufficiently great to prevent degradation, yet not great enough to cause general agradation; that along a mountain base the rock planes will have the form of coalescing fans apexing at canyon mouths; that rock planes and rock fans are repeatedly regraded at lower and lower levels, due to decrease in load of streams and to other causes; and that such regrading leaves the mountain front bordered by terraces or benches at various levels, each terrace being a high-level remnant at the inner border of a formerly more extensive pediment or rock fan.

Terraces of the type just described have been examined by the writer along the bases of Rocky Mountain ranges in Colorado, Wyoming, and Montana; among the ranges of western Texas, central and southern New Mexico, southern Arizona, and southern California; northern Africa, South Africa, and in both the north and south islands of New Zealand. It should be anticipated that coastal portions of California, northern Africa, South Africa, and New Zealand might exhibit these forms, and that when bordering the sea they mght be confused with terraces of marine origin.

Examination of the literature suggests that in each of the four localities last mentioned there may be terraces open to either the fluvial or marine interpretation. In California Warren D. Smith has described terraces, interpreted as of marine origin, from areas described by Lawson as exhibiting no marine forms. In northern Africa some of the classic marine terraces of De La Mothe have later been interpreted by Baulig as fluvial forms. In South Africa Krige has interpreted as of marine origin a series of planed rock surfaces covered with gravels devoid of marine fossils. Planes of lateral stream corrasion are beautifully developed in this region both inland between the mountain ranges and on the coastal borders. Schwarz interpreted some of the coastal terraces first as of marine origin, later as fluvial forms, finally returning to the marine interpretation. In New Zealand Cotton has ascribed to wave action a number of coastal terraces, some of which the present writer is inclined to regard as graded surfaces of subaerial origin.

The most remarkable coastal terraces it has been the writer's privilege to see are found along the northeastern coast of the South Island of New Zealand. These have been studied extensively by Jobberns, in part in collaboration with King, Sylvester, and others. Jobberns found that beveled rock surfaces or platforms are developed chiefly on weak rocks, flanking the hard-rock ranges; that they may be largely or wholly missing from hard-rock portions of the coast; that normally they are cut back but slightly into the hard-rock cores of the ranges; that the inner edges of the platforms are sometimes nearly or quite devoid of covering deposits, although

farther out a cover many feet thick may exist; that elsewhere the inner portion of the cover is very thick and has the form of coalescing alluvial fans; that this cover is usually devoid of fossils and often demonstrably consists wholly of stream-borne alluvium; that at its base, resting directly on the beveled surface of the tilted underlying rocks, is sometimes found a layer of well-rounded, obviously water worn gravel, free from fossils; that the beveled rock surface on which the covering deposits rest is remarkably plane for long distances; and that some of the lower platforms are in general apt to be more extensive and more clearly defined, the higher ones more restricted in area and less perfectly preserved.

If one will examine the literature describing pediments and rock fans, he will find that every one of the phenomena catalogued above and reported by Jobberns for his coastal platforms is typical of the rock planes described by Gilbert, McGee, Paige Lawson, Bryan, Davis, Johnson, and others. Jobberns was frankly puzzled by many of the features he observed, because they did not correspond to his expectations based on the marine theory of origin. The absence of a cover at the inner edge of the rock platform he found "peculiar" and "curious"; in the general absence of marine fossils, and in the abundance of undoubted stream alluvium resting directly on a smoothly planed rock surface, he found "a difficult problem," and he "considered in the field the possibility of the platform having been cut and covered entirely by the action of the streams." He concluded that "solution of the problem must be left for future work." This investigator clearly recognized the fact that discrimination of marine from fluvial forms can be extremely difficult. Certain terraces, which in his earlier writings he regarded as undoubtedly marine, he later interpreted as fluvial in origin. Others he ascribed to a marine origin on evidence which does not always seem conclusive. In a few cases the former presence of the sea on a beveled rock surface could clearly be demonstrated by the presence of marine shells and borings, beach ridges, or other evidence. But in the vast majority of cases direct evidence of the intervention of the sea was lacking.

Details of evidence presented by the various investigators mentioned above, and whether these investigators were right or wrong in the their conclusions, are not relevant to the present discussion. We are here concerned only with the fact, not yet adequately appreciated, that there do exist marine and fluvial forms which are so similar in appearance that their discrimination is difficult. There are, in fact, various types of terraces of nonmarine origin which may easily be mistaken as records of former sea levels. But enough has been written to establish the thesis of this part of our discussion: that the marine origin of terraces bordering the sea cannot safely be taken for granted.

CRITERIA OF ORIGIN

To be safe, any conclusion involving a marine history must be based on specific study of the question of origin. When dealing with rock platforms, for example, there must be, among other things, a critical analysis of the characteristic features of wave-abrasion surfaces and stream-carved rock planes. Before one can discriminate between the two, he must know the degree of slope of each at outer and inner margins, nature of the transverse profile of each, peculiarities of surface contours of each, and the nature of each as seen in stream-cut or wave-cut cliffs parallel, oblique, and at right angles to bordering highlands. Adequate field studies of these

elements have not yet been made. Such studies may show that the two types of surfaces have many things in common but also have a few distinguishing peculiarities such as, for example, the more nearly plane surface of the inner border of a marine platform as contrasted with the undulating inner border of a pediment. These distinguishing peculiarities will be the critical points of evidence for which search must be made.

The nature of the covering beds will be less critical, for the sea has undoubtedly during its oscillations of level, especially those connected with glaciation and deglaciation, as well as during recent oscillations of land level, advanced over stream-carved platforms bordering the sea and left marine gravels and shells on subaerially formed surfaces. Streams have just as surely spread their alluvium over marine platforms from which the sea has withdrawn. If, however, it be desired to determine the recent presence of streams or sea on the platforms, the distinguishing features of fluvial and marine gravel deposits of different types should carefully be formulated. These may be found not in the forms of component pebbles, since flat, rounded, and other-shaped pebbles are found in both marine and fluvial beds, but rather in the presence of material derivable or not derivable from adjacent streams; material existing or not existing at points along the coast from which beach drifting or other marine transport is known to take place; fresh-water or marine shells of recent date and not derivable from beds reworked by streams or waves; imbrication of pebbles characteristic of flood plains, alluvial fans, beach ridges, and other fluvial or marine deposits; and presence of floodplain bars and channels, meander scars and scarps, beach ridges and swales, beach cusps, and other fluvial and marine forms. All discriminating criteria should be formulated not in the inner consciousness of the observer, on the basis of what he thinks streams and waves ought to do, but rather on the basis of a critical study of undoubted fluvial and marine forms and deposits in the field. Such study should be made in the light of all information the investigator can derive from the field observations of other students as recorded in the literature; but he should not be unduly influenced by the conclusions of those other students, especially where it appears that their conclusions are based on inadequate field study, or on theoretical deductions not adequately tested.

Once the discriminating criteria are carefully established, and their occurrence in the field demonstrated, conclusions should not extend further than the criteria justify. If critical evidence proves that a platform is cut by the sea, it is not permissible to conclude that the cover resting on the platform is also of marine origin. If presence of marine shells and other criteria demonstrate that a cover resting on a rock platform was laid down by the sea, it is equally dangerous to conclude that the platform itself was also cut by the sea. In the first case the reasoning is faulty, because it is known that streams in the recent past have been advancing over all sorts of forms produced by the sea. In the second case the reasoning is faulty, because it is known that the sea has recently advanced over all sorts of forms produced by stream action, later retreating from some of these, but still covering others; and that certain of these oscillations of sea level were so rapid that no extensive benches or cliffs could be cut, although marine gravels, sands, and clays, containing marine shells, were laid down.

It must further be recognized that the critical evidence essential to the solution of a given problem may not be obtainable. Where rock platforms of whatever origin have been deeply buried under later deposits, and subsequently revealed to present-day observation at relatively few points along sea cliffs and valley walls, those points perhaps being ones where critical evidence is not available, it may be quite impossible to say what agency carved the platform. In this case there is only one scientific procedure: to report the facts, to explain the reasons for the inadequacy of the evidence, and to decline formulation of a final conclusion so long as evidence on which to base such conclusion is not available. Not infrequently the only safe conclusion is that reached by a French investigator of the cause of exfoliation, who wrote at the end of his elaborate discussion: "En somme, la question reste ouverte."

DETERMINING ALTITUDES OF TERRACES

DIFFICULTIES

To determine the altitude of coastal terraces and river terraces as a basis for correlation is a task far more exacting than it appears. In papers cited near the beginning of this article the writer has set forth some of the common defects which most frequently vitiate terrace-altitude figures, such as failure to record what parts of terraces the figures refer to, recording altitudes for portions of terraces which bear no fixed relation to former water planes, and using barometric observations which are too inaccurate for the purpose in view. Here emphasis is laid upon certain difficulties, some of them not stressed before, which are repeatedly encountered when seeking to verify in the field conclusions set forth in published reports.

USE OF THE BAROMETER

Because the barometer is a convenient and much used instrument in terrace work, a further word concerning its limitations seems desirable. That the instrument has value for correlation work under certain conditions is fully recognized. If the observer finds on a given coast or river valley (1) that he has to deal with but a few well-developed terraces, (2) that these are separated one from the other by vertical intervals amounting to some scores or hundreds of feet, within which intervals terrace remnants are absent, (3) and that he can pass from sea level or stream level to the several terraces without much lapse of time or great change in atmospheric conditions, and can check the field readings against a stationary instrument read at frequent intervals through the period of field observations, then he may have little difficulty in securing data which will enable him to correlate separate portions of the same terrace with each other and so determine which terraces, if any, are approximately horizontal, slightly inclined, or strongly tilted. But experience shows that barometric observations are not sufficiently accurate or sufficiently consistent to be trusted for terrace-correlation work where the vertical spacing between terraces is measured in tens of feet instead of in scores or hundreds; where the intervals between major terraces show remnants of intermediate terraces which might be confused with those subject to correlation; where much time must elapse between the barometric observations to be compared, or where areas to be compared are far apart and possibly subject to strongly contrasted atmospheric conditions; where it is not practicable to check field readings against a control instrument located within the area

DETROIT PUBLIC LIBRARY

being studied; or where it is important to determine either the absolute horizontality or the exact degree of inclination of terraces. It is conceded that barometric observations, even under these last conditions, might happen to be correct, but they are not sufficiently dependable to be made the basis of so difficult and delicate an operation as the critical correlation of former marine and stream levels. Employment of some surveying instrument of reasonable precision alone will give dependable results.

SIGNIFICANT PARTS OF RIVER TERRACES

It should not be necessary to emphasize the fact that in correlation studies of either marine or fluvial terraces the only altitude figures which can properly be used are those determined for such significant parts of terraces as bear a known and fairly constant relation to former water planes. Unfortunately, very elaborate studies of both types of terraces continue to be made in which this fundamental fact is ignored. Stream terraces (Pl. 1, fig. 1) often consist of a planed rock surface upon which lies a veneer of stream gravels covered in turn by a deposit of slope wash from the valley walls, talus from the same source, alluvium brought in by lateral tributaries, loess, dune sand, or other extraneous material. The deposits overlying the gravels laid down by the main stream are often tens or scores of feet in thickness and not infrequently exceed 100 feet. In such cases there is no element in the terrace *form* which is significant for correlation purposes. The altitude of the front edge of the terrace (Pl. 1, fig. 2, F) is purely accidental, depending on the extent to which the terrace has been undercut by lateral planation of the stream at a lower level and on the thickness of the extraneous deposits (slope wash, etc.) at that point. The altitude of the rear of the terrace (R) is determined by the irrelevant factor of slope wash or talus accumulation. The only surfaces whose altitudes can be employed in correlation studies are normally altogether concealed under an unknown thickness of slope wash and other debris, or badly obscured by slumping on the frontal scarp of the terrace. Such significant surfaces are the upper level of the veneer of stream gravel (G) and the beveled rock plane (P) on which it rests. Both may occasionally be observed where recent stream undercutting along the frontal scarp gives fresh exposures, or in the walls of side ravines which trench the terrace. These are the critical points which may fortunately reward one's search; and it is at such points only that worthwhile altitude determinations can be made. Elevations taken on the cover of extraneous debris are seriously misleading, since the thickness of the cover is not constant, and figures relating to its variable surface altitude introduce errors of unknown but often great magnitude.

What we observe where a terrace consists of a planed rock surface veneered with gravel and the veneer deeply buried under slope wash and other debris should put us on our guard when dealing with terraces cut wholly in earlier stream deposits. Here the distinction between surface of planation, gravel veneer, and extraneous cover may be difficult or even impossible to make. Yet altitude determinations based on the surface of the terrace form will be just as misleading here as they are when the errors are made obvious by the favorable exposures occasionally found under conditions described above. If the surface of the terrace is obviously the old flood plain of the master stream, with its bars and channels, meander scars, and

FIGURE 1. POWELL TERRACE OF THE SHOSHONE RIVER
(F) Front edge of terrace form; (W) slope wash deposit, over 100 feet thick in places, constituting nearly half of the terrace structure; (G) upper level of river-deposited black volcanic gravels. The gravel sheet is thin and rests on a river-planed bedrock surface which is concealed by gravel and other debris washed far down the slope.

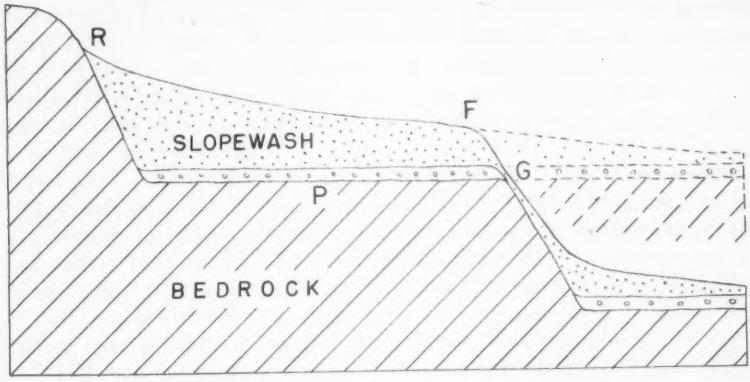


FIGURE 2. CROSS SECTION OF TYPICAL SHOSHONE RIVER TERRACE
(R) Rear of terrace; (F) front edge of terrace; (P) beveled rock plane; (G) stream-gravel veneer.

SHOSHONE RIVER TERRACES, WYOMING

other determinable features of diagnostic value, little difficulty is to be anticipated. But, where the terrace surface shows that perceptible downward slope toward the axis of the valley which is characteristic of slope wash, the coalescing fan forms of alluvium poured out by lateral streams, or other evidences that an extraneous cover has been deposited on the former valley floor, the investigator must abandon his measurements of altitudes until critical studies of all available exposures enable him to place with confidence the level of the true valley floor beneath its extraneous cover. Should this last prove impossible, terrace-correlation studies at that locality may better be abandoned. It is better to have no result than to have a misleading result.

SIGNIFICANT PARTS OF COASTAL TERRACES

The difficulties involved in the study of coastal terraces are often much greater than those encountered in river-terrace studies. Platforms of marine abrasion, like planes of lateral stream corrasion, are apt to be buried under a thick and variable mantle of slope wash and other debris. But, whereas almost any part of a stream-carved platform, when exposed to observation, may enable the investigator to secure altitude data of critical value, it is only the altitude of the inner margin of a wave-cut platform which has any significance. This is because platforms of marine abrasion slope seaward indefinitely and at variable angles, and the only part which even approximately fixes a former level of the sea is the inner edge, at the base of the wave-cut cliff. Unfortunately, it is usually portions of the platform far out in front of the ancient cliff which are best exposed to view, most frequently where waves of a lower sea level are cutting into the outer edge of the elevated bench. The critically important inner edge is most commonly concealed under the cover of extraneous debris; and it is precisely here that the cover is apt to have its greatest and most variable thickness. The investigator is fortunate if he can find a sufficient number of exposures, either where the debris cover is lacking or in the walls of valleys cut through the debris and into bedrock below, to fix the altitude of the planed rock surface at its inner margin. If he can do this in a large number of places, reliable correlation may be practicable.

The possibility of determining indirectly the altitude of a marine-abrasion platform's concealed inner edge deserves attention. If the altitude of the platform is determinable a short distance out from its inner margin, one may suppose, providing earth movements have not distorted the platform surface, that the inner margin is somewhat higher. If the slope of the platform at various distances from the inner margin has been determined from other exposures, one may even compute the *probable approximate altitude* of the inner margin at the point in question; or he may compute the limiting altitudes above which and below which the surface is reasonably sure not to be found. But the value of indirect determinations of platform altitudes is very limited and falls rapidly with increase of distance from cliff base to the exposures used as a basis of calculation. The reasons for this will appear when we discuss, in a later section, the importance of proving identity of altitudes or regularity of vertical spacing as a basis for terrace correlation.

If, with the foregoing discussion in mind, one turns to the literature of coastal-

terrace correlation, he will be surprised to discover how frequently altitude records fail to provide a sound basis for correlation purposes because the altitudes determined relate to features having no significance. Some writers record altitudes for the front margins of terraces, sometimes based on exposures of the planed rock surface, sometimes based on the alluvial cover. In both cases the figures may bear no significant relation to former water planes. The altitudes of flat-topped hills, believed to be remnants of a badly dissected marine platform, are recorded, even when the positions of such surfaces with respect to the supposed former sea cliff (now destroyed by erosion) is wholly unknown. Countless observations are based on the present surface of the terrace *form*, where this is determined by a cover of slope wash or other debris of great thickness.

We need hardly stress the fact that all such altitude figures are wholly unsuitable for correlation purposes. They falsify any correlation attempted and lead the investigator himself and all who follow him to erroneous conclusions. In one case a planed rock surface approximately 150 feet above sea level was correlated with an alluvial surface at 160 feet, although the alluvium was very thick, and the planed rock surface under it was far below the 150-foot level. In another instance two terraces were correlated because the surface of each was approximately 250 feet above sea level at its inner edge. But in one case the 250-foot elevation was taken on an alluvial cover known to be of great thickness, and diagrammatically represented by the author as approximately 100 feet thick at its inner edge. The result was to throw into the 250-foot terrace group a terrace which should have been assigned to the author's 150-foot group. Such instances could be mutliplied indefinitely. We have gone far enough, however, to demonstrate that to determine altitudes of terrace forms, without critical inspection of terrace structures and without discovering the only features for which altitude determinations can have any significance, is to assure in advance the failure of terrace correlation to achieve results of any real value.

CORRELATING TERRACE ALTITUDES

LOGICAL BASIS OF CORRELATION BY ALTITUDE

Once terrace altitudes are properly determined, we pass to correlation of the terraces on the basis of altitude figures. Here again, as experience abundantly demonstrates, there is ample opportunity for the student of terrace correlation to find himself involved in serious difficulties.

Let us first get clearly in mind just what is meant by "terrace correlation on the basis of altitude figures." It is not, in any sense, a method of determining the approximate altitudes of terrace remnants known to belong together. It is not a method of determining whether such terraces are horizontal or inclined. Very rough altitude determinations would suffice for the purposes just named; and it is unfortunate that many students have employed, in terrace-correlation studies, field methods appropriate only to these latter purposes.

"Terrace correlation on the basis of altitude figures" is a special method of determining *which terrace remnants along a valley wall or coastal border formerly constituted*

parts of the same terrace. It is a method very frequently employed because other methods cannot be used. Far safer, and much easier to apply, is the method of *continuity of form.* If the terrace form can be traced from point to point, with gaps so small as to leave no possibility of doubt that the same terrace is picked up after each short hiatus, the student can safely identify the several members of a terrace series at different points. He can then proceed with confidence to determine, roughly or with precision, as he may elect, whether the terraces are horizontal or inclined. This is often possible in the case of the lower and better-preserved terraces bordering a river.

Unfortunately, terraces were not always formed continuously along valley wall or coastal border; and subsequent erosion has often destroyed much of what was formed. Such is usually the case with the higher older terraces of a stream valley or coastal border, while even the lowest coastal terraces are frequently discontinuous. Under these circumstances the method of continuity of form cannot be used. Correlation on the basis of *identity of fossil contents* cannot often be employed (1) because the physical conditions in terrace deposits are commonly unfavorable to the preservation of fossils and (2) because successive terraces are commonly formed in periods of time too short to permit marked differentiation of faunas either through evolution of the same faunas *in situ* or through migration of other faunas into the area. Correlations on the basis of *degree of terrace dissection*, on the *degree of decomposition of gravels*, and on the *depth of weathering of soils* are frequently untrustworthy, due to the fact that differences in distance from major water courses and variations in texture of stream dissection, in position of ground-water level, in climatic conditions, and in other factors often cause greater differences in the supposed criteria named than does the amount of time elapsed since the terraces were cut. Difficult though it be, correlation on the basis of altitude figures may offer the investigator his only chance of solving a given terrace problem.

It is highly important to appreciate the fundamental difference in purpose and methods between *determining the altitudes* of various terrace remnants known to have been formerly parts of the same single terrace and *proving by altitude determinations* that various terrace remnants were parts of the same terrace. In the first case the principal facts to be determined are the amount of change in level of land or sea or stream, and whether the elevated terrace is horizontal or inclined. These facts may appear even if the measurements are crude and approximate; indeed, they will appear in sufficient measure for some purposes if half the altitude figures be of doubtful validity or seriously in error, since a reasonable number of correct determinations will establish the general situation.

In the second case it is not merely the height above stream or sea and the general altitude of a terrace which is to be determined; the very existence of the terrace itself, the thing to be measured, is at issue. Are the various terrace remnants merely records at a multitude of diverse and insignificant points of an entrenching stream or a falling sea level, which left now here, now there, in its downward progress, evidence of its presence where local conditions favored? Or are they parts of one single terrace marking a significant pause of the stream, or a significant stand of sea level, at a given horizon? Or are they parts of two or more such terraces, marking two or

more significant horizons? We are no longer trying to measure a thing. *We are trying to prove the existence of the thing to be measured.* When this is our object, crude and approximate altitude determinations will not suffice, and a few figures of doubtful validity may easily vitiate the conclusions reached.

NECESSITY OF PRECISE ALTITUDE FIGURES

To prove, by altitude measurements, that various terrace remnants are parts of the same original terrace, marking a single significant stand of the sea or position of stream level, the figures must not only relate to the same significant parts of the terrace, as explained in the preceding section of this paper, but they must be reasonably exact. This fact depends upon the following principle:

Several terrace remnants on different parts of a coast or on different coasts, or on different parts of a valley wall, even when of entirely different age and different origin and hence absolutely uncorrelatable, may by chance be somewhere near the same altitude above sea or stream. But they are not likely by chance to be at exactly the same altitude. It is possible, but not very probable, that they may by chance be very close to the same altitude. *Identity of altitude*, therefore, is ordinarily the only proof of identity in age and origin of terraces; when this is satisfactorily established over a long distance it may create a strong presumption in favor of a uniform uplift of the land, a drop of sea level, or other conclusions of equal import. *Close similarity of altitude* affords evidence of a lower order of reliability. *Approximate similarity of altitude*, since it may well be fortuitous, affords no evidence of value and should not seriously be considered in any scientific study of terrace correlation.

It is evident that the expressions "identity of altitude," "close similarity of altitude," and "approximate similarity of altitude" require definition and that these definitions will vary according as the vertical spacing between the different terraces of a series be small or great. "Identity of altitude" does not mean that three terrace remnants, in order to be considered as having identital altitudes, must each be precisely 151.078 feet above a given datum plane. Nature does not fashion terraces or other forms with such precision. If the measurements are 153 feet, 148 feet, and 150 feet, respectively, the terraces may reasonably be said to have identical altitudes.

How far the altitude measurements may depart from the mean, and still be considered substantially "identical," will depend upon the vertical spacing of the terraces. If one is trying to ascertain whether a variety of scattered terrace remnants should be correlated with members of a supposed series of three terraces occurring at altitudes of 20 feet, 40 feet, and 60 feet, respectively, it is obvious that terrace remnants found 19 feet and 22 feet may with reasonable safety be regarded as identical in altitude with the 20-foot terrace, remnants at 38 feet and 41 feet as identical with the 40-foot terrace, and so on. Terrace remnants found at 24 feet and 36 feet may be regarded as closely similar in altitude, the first to the 20-foot terrace, the second to the 40-foot terrace. But terrace remnants at elevations of 28 feet and 31 feet certainly cannot be considered as identical, or even closely similar, in altitude with any of the standard series. When due allowance is made for original variations in terrace altitude and subsequent changes due to erosion and weathering,

not only on the remnants thus measured but also on the terraces with which they are compared, it is evident that both the 28-foot and the 31-foot levels may belong with either the 20-foot terrace or the 40-foot terrace; or may represent a distinct terrace at an intermediate level. It is folly to attempt correlation under such circumstances.

On the other hand, if well-developed terrace remnants are found at altitudes of 120 feet and 240 feet, and not between those levels, other strongly developed remnants found at 124 feet and at 236 feet may be regarded as substantially identical in altitude with the 120-foot and 240-foot terraces respectively; while remnants at 128 feet and 231 feet are so closely similar to the 120-foot and 240-foot levels that correlation with terraces at these levels is reasonably safe. The range of 8 or 9 feet from the nearest standard figures which is fatal to correlation when terraces are closely spaced is not fatal when terraces are widely spaced.

The foregoing principles have been stated with respect to terraces supposed to be substantially parallel to the datum plane (*e.g.*, sea level or lake level) from which measurements are made. Where warped coastal terraces are involved, the same principles apply, but in a modified form, to take account of progressive changes in terrace altitude.

Where river terraces are involved, and elevations above present stream level are being measured rather than elevations above sea level or lake level, other principles enter. It is not safe to assume, as has been done in some published studies, that terrace systems resulting from successive rejuvenations of a stream, whether by successive uplifts of the land or by successive drops of sea level, will maintain uniform altitudes above the stream throughout its course. The assumption is based on the supposition that at each rejuvenation the stream will cut downward until in its new position it is everywhere at the same depth below its former valley floor, at least throughout that portion of the stream's course where terracing is in progress. It is thus assumed that a given terrace, sloping down valley in harmony with the stream's descent, should be everywhere, let us say, 90–100 feet above present stream level, a second higher terrace everywhere 260–270 feet above present stream level, and so on.

But it seems most improbable that during a succession of incisions, extending over a period of time sufficiently long to permit repeated down cuttings and repeated valley widenings throughout a vertical distance of hundreds of feet, a stream will return to the same profile of equilibrium after each incision. It is more likely that differences in the quantity of debris fed into the stream at higher and at lower levels, changes in climate, and other variable factors will cause the stream to grade its profile sometimes at a steeper, sometimes at a gentler slope. In such case terrace remnants of an older and higher flood plain or valley floor will, when followed down stream, be found either progressively closer and closer to present stream level or higher and higher above present stream level, according to whether the present stream profile is more gently sloping or more steeply sloping than its earlier profile. In either of these cases altitudes of correlatable terrace remnants will not, over any considerable distance, be even approximately uniform. They will progressively increase or decrease at a rate which may itself progressively change.

If with these principles in mind one turns to the literature of terrace correlation,

he will be interested to note that terraces are, for example, described as occurring at 150 feet, at 250 feet, and at 350 feet above sea level. Since it is in the highest degree improbable that Nature would space land uplifts or sea-level depressions in such manner that the intervals between them are not only an even 100 feet, but that the elevations fall exactly on the half-hundreds, the figures at once advertise themselves as being not real measurements of altitude but generalizations of, or inferences based on, such measurements. A study on the Atlantic coast reports terraces at 100-foot intervals, but in this case the terraces are said to occur at elevations of 100, 200, 300 feet, etc., instead of 150, 250, 350 feet. A French investigator has reported successive terraces at intervals of 100 meters. We must conclude either that a wise Providence not only pays remarkable attention to the magic number 100, but also nicely adjusts uplifts of the land and lowering of sea level to the particular system of measure prevailing in each country; or else that actual elevations of very different character are by observers so roughly approximated to the nearest even figure as to make them valueless for correlation purposes. The latter interpretation seems the more reasonable.

Generalized or inferred figures have no place in a correlation study, and conclusions based on them must always be regarded as untrustworthy.

CORRELATION OF ADJUSTED VALUES

It should not be difficult to understand why precise actual measurements of really significant elements of terraces can alone be used in correlation studies. Any other procedure destroys the very basis of proof that terrace remnants are to be correlated. This is because it is the *closeness of the agreement* between altitude figures, a degree of agreement too close to be explained by chance, which forms the sole evidence that terrace remnants really do belong together. Only when altitude figures are exact and agreement between them is close can one safely conclude that terrace remnants represent fragments of a single platform carved during the same stillstand of land or sea, or parts of the same former valley floor of a stream. The moment that one begins to assemble a variety of altitude measurements into a few groups of generalized figures, and to correlate these generalized figures with one another, the possibility of securing trustworthy results vanishes.

When direct comparison of actual altitude figures is abandoned, and generalized figures are compared, there almost inevitably enters into the picture a procedure which the writer has called "the correlation of adjusted values." Because this procedure is based on reasoning which to many seems sound, and is frequently encountered not only in terrace correlation but also in other correlation studies, it seems necessary to examine it somewhat critically. And because the fallacy involved in the procedure will the more clearly be evident if we take an illustration completely outside the field of science, let us have recourse to an example encountered in one of our great English classics.

In his famous diary Samuel Pepys points out that after the Great Fire in London one Griffin told him that "it was observed, and was true," that the fire burned just as many parish churches as there were hours from the beginning to the end of the fire;

and, further, that there were just as many churches left standing as there were taverns left standing in the part of the city not burned, "being, I think he told me, thirteen in all of each; which is pretty to observe."

Now if the accordance in figures was really so remarkable as Pepys' informant stated, the famous chronicler was fully justified in considering the fact something "pretty to observe." But was the accordance really so remarkable? That depends on how the figures were obtained. The beginning of a fire may be fixed with some precision. But when does the burning of a great city end? Hours, or days, or perhaps even weeks after the beginning, according to what a given individual chooses to interpret as the end. Even the number of churches is subject to individual interpretation. Likewise the number of taverns. Let us suppose that the fire left standing 10 churches at that period in active use, 2 church buildings previously sold and appropriated to secular use, and 2 churches in process of construction but not yet occupied. Let us suppose further that the fire also left standing 19 taverns, 13 being of the type providing meals and lodgings for travelers, the remaining 6 being public houses selling liquors only. There is nothing significant in these figures as stated. But one seeking to discover a hidden relationship of vital import may easily do so. He finds that if he counts all the completed church buildings, including the two sold for secular use, but excludes the two uncompleted churches, the total is 12. Similarly, counting all the taverns providing meals and lodgings, but excluding those which are merely "pubs," the total is 13. This is very close to the figure for churches. Now further examination shows that one of the uncompleted churches was so nearly finished that it was dedicated the Sunday following the beginning of the fire. It seems wholly justifiable, therefore, to include it in the list of preserved churches. The figures now stand 13 churches left standing to 13 taverns left standing. The correlation is perfect. Such exact agreement in figures, recounted to one unfamiliar with the manner of determining it, may well seem "pretty to observe."

In the case pictured above the correlator of churches with taverns may have been quite unconscious of the fact that he fabricated in his own mind the basis of the precise correlation which seemed to him so significant. Let it be noted, furthermore, that every step he took was in itself wholly reasonable and justifiable. He did not conceal or distort the facts regarding any church or tavern, and his reasons for including some and omitting others among them are plainly stated. The only thing which is unreasonable and unjustifiable is to attach any significance to a coincidence of figures obtained by this procedure.

It is very easy, in terrace-correlation studies, for the investigator quite as unconsciously to make the same capital error. Terraces when first fashioned, let us say by the waves of the ocean, do not have their inner margins absolutely horizontal. Wave cutting takes place at slightly higher levels in some localities, at slightly lower levels in others. Any one figure for the elevation of the terrace is at best, therefore, an average figure. The same erosion which completely destroys parts of the original terrace may somewhat reduce other parts below the normal level. It accordingly seems legitimate to add something to an abnormally low figure to bring it up to the standard elevation. Slope wash and other debris may be observed to form the terrace surface where the measured altitudes seem abnormally high. Of course

something should be deducted from such altitude figures to bring them closer to the true elevation of the wave-cut platform. The elevations of beveled rock surfaces found on isolated hill tops, evidently far out from the base of the ancient sea cliffs, may require quite a considerable addition before they will be even approximately near the true position of the ancient water plane. Where altitudes are determined by barometer, the eccentricities of which instrument are well known, there is reason to suppose that noticeable departures from the normal, not otherwise explained, may be due to the limitations of the instrument employed. In any case, generalized rather than precise figures are what we must expect when the barometer is the basis of operations.

When the obviously appropriate corrections are made it is found, let us suppose, that most of the terrace remnants along a given coast fall into two groups—one having altitudes ranging from 170 to 235 feet, the average altitude being about 200 feet, the other having altitudes ranging from 275 to 335 feet, with an average of 300 feet. There remain a limited number of intermediate terrace elevations so far outside these limits that no reasonable correction applied to them would bring them into harmony with the great bulk of the observations. It is manifest, however, that the sea must have fallen from the 300-foot level to the 200-foot level; and it is only reasonable to suppose that in doing so it left occasional traces of its presence at intermediate levels where local conditions favored the rapid development of shore forms.

The evidence thus points to two major terraces at 300 and 200 feet respectively above sea level, with less well developed traces of the sea at intermediate levels. Now the same or another investigator has found, by similar studies, a terrace at about 300 feet and another at about 200 feet on the coast of Erewhon, several thousand miles to the south. This remarkably close correlation in altitudes over so great a distance precludes uplift of the land as the cause of present terrace altitude. Eustatic drop of sea level offers the best explanation of the phenomenon.

The reader can duplicate every phase of the foregoing hypothetical terrace correlation from actual examples described in the literature. Each step indicated above seems logical and proper and in most cases would be, were the object in view to get the best approximate altitude figures for a terrace known to exist. But when the object in view is to prove whether terrace remnants represent one, or two, or several distinct terraces, formerly more or less continuous and representing as many distinct and significantly long stands of the sea, it must appear that the procedure as a whole is identical in character with that involved in the correlation of churches with taverns. The real basis of correlation has not been found in Nature. It has unconsciously been fabricated in the mind of the observer, by a series of steps each one of which in itself seems wholly justifiable. But taken together these steps lead imperceptibly first toward substitution of inferred or generalized figures for those actually determined in the field; next to segregation of these inferred and generalized figures into groups; then to substitution of averages or standard figures for the more widely ranging figures of these groups; and finally to correlation and conclusions based on the standardized figures, now several times removed from, and bearing slight relation to, the actual figures determined in the field. *The sole proof of correct correlation being identity of altitude figures,* that proof is destroyed when the altitude figures are

fictitious rather than real, and when the identity is manufactured rather than natural. The things correlated have adjusted values instead of their initial values; and it is the process of adjustment which makes the correlation seem close.

LIMITATIONS OF THE METHOD OF CORRELATION BY ALTITUDE

Many correlations of terrace remnants must be regarded as of doubtful validity because the investigators correlated ranges of altitudes rather than individual observed altitudes (or possible ranges of water planes[2] determined independently for each individual point) and employed a range of altitudes nearly as great as, or even greater than, the interaltitude range. In the illustration given above one group of terrace remnants is supposed to range in altitude from 170 feet to 235 feet above sea level, the second from 275 feet to 335 feet. Although the two terraces supposedly represented are only 100 feet apart, the vertical range of the first group of altitudes is 65 feet, of the second 60 feet. The interval between the two groups, or interaltitude range, is only 40 feet. Were terrace remnants by chance distributed equally throughout the entire vertical range, there would necessarily be more remnants within each supposed terrace range than in the intervals between them, simply because the scope of the former is greater. This apparent concentration of terrace remnants within the terrace range is wholly fictitious, for by assumption the remnants are equally distributed.

But terrace remnants will not normally be distributed equally. Their chance development at some places and not at others during a steady fall of sea level, and their preservation at some places and not at others under the vicissitudes of subsequent erosion, will be more likely to give a purely fortuitous "bunching" of remnants in certain zones and not in others. Where irregularly spaced hill tops of nonmarine forms are mistaken for marine-terrace remnants, their altitudes will, for a variety of reasons quite unrelated to the existence of supposed former water planes, be found more numerous at some levels, less numerous at others. If the observer, in selecting the zones of abundant figures, makes the ranges sufficiently great, he can bring within them most of the altitudes measured, leaving perhaps but a few scattered observations in the narrower intervening zones. Here a purely fortuitous concentration is so accentuated by the device of enlarging the zones in which such concentration occurs and reducing the zones in which records are more sparse that a grouping of figures which may have no significance whatever begins to take on the color of real evidence.

Of course, if there were originally two or more continuous terraces, and if these terraces were warped so that each extended upward and downward through a vertical zone of considerable extent, and subsequent erosion then left scattered remnants at various levels; or if measurements were taken sometimes on debris burying the original terrace structure, and sometimes on parts of the terraces worn below the original level; then the altitude figures would have precisely the distribution described above. The fact that remnants of formerly continuous terraces will, under conditions easily conceived, give such groupings of altitude figures has doubtless

[2] Johnson, Douglas (1933) *The correlation of ancient marine levels*, Cong. Inter. Géog., Paris, 1931, C. R., Tome II, p. 42–54.

led many to accept such groupings as proof of the existence of former terraces. The fallacy of such reasoning lies in failure to realize this vital fact: Altitude figures fully explicable as the product of warping, erosion, and aggradation of terraces definitely known on independent evidence to exist cannot be used to prove the reality of terraces the existence of which is reasonably open to question.

Correlation of terraces is frequently invalidated because the investigator thinks in terms of the *absolute distance* which separates one group of altitude figures from another group, when he should think in terms of *relative distance.* A problematical terrace remnant with an altitude of 225 feet is fairly close to a group of similar remnants in which the altitude range is from 240 to 250 feet, if one pays attention only to absolute figures; for 15 feet, or even 25 feet, is not a great distance, when one considers the inevitable irregularities of terrace surfaces. In a word, the accordance seems fairly close. But if the altitude range of the next lower terrace is 200 to 210 feet, then the figure 225 feet is as remote from 240–250 feet as it is possible for it to be without falling even closer to the next lower group. Instead of fairly close accordance, we have, in fact, the maximum possible discordance. Correlation on the basis of altitudes is wholly impossible, for the observer has no means of knowing whether the remnant in question belongs with the remnants 15 to 25 feet higher or with those 15 to 25 feet lower.

The following correlation table is taken from one of a number of papers in which authors have attempted to correlate terraces in different parts of the world with the famous terraces of the Mediterranean coast. In the particular paper selected terraces on the coast of *X*, thousands of miles from the Mediterranean, are correlated with type terraces of the latter locality. For convenience of reference I have labeled the supposedly corresponding terraces a a′, b b′, c c′, and so on.

	Coast *X* Feet		Mediterranean Coast Feet
a	40–60	*a′*	60–66
b	120–150	*b′*	92–98
c	230–250	*c′*	181–198
d	330–380	*d′*	295–328
e	500–525	*e′*	486
f	650–700	*f′*	670
g	800–900	*g′*	870
h	1000–1200 (estimated)	*h′*	1075

Figures in the second column have a false appearance of exactness, since, as the author fully explains, they represent conversion into feet of altitudes originally given in meters. When we examine the column of figures for Coast *X* we note first that the ranges of altitudes for the several terraces are often relatively large in comparison with the vertical distances between the terrace altitudes. Thus the range for terrace *d* is 50 feet when the distance to the next lower terrace range is only 80 feet; the range for terrace *f* is 50 feet, while the vertical distance to the next lower terrace is 125 feet, to the next upper terrace but 100 feet; the range for terrace *g* is 100 feet, when the distance to the next lower terrance and to the next upper terrace is in both cases only 100 feet. The estimated range for terrace *h* is 200 feet, or double the distance (100 feet) to the next lower terrace.

Examination of the text shows that even these wide ranges represent inferred, generalized, or standardized figures. Actual field observations range far more widely, and in the text even the generalized figures often differ significantly from those used in the correlation table. Thus terrace *a* (40–60 feet) is elsewhere referred to as the 50–60-foot level, while a level 30–40 feet is also recognized. Terrace *b* (120–150 feet) is usually referred to in the text as "150–160 feet." Terrace *c* (230–250 feet) is more frequently described as the terrace "at 250 feet," but the actual elevations cited range from 220 to 290 feet, a range of 70 feet when the distance from the highest measurement on this terrace to the lowest on the terrace next above is only 10 feet. Terrace *d* (330–380 feet) is also referred to as "300–350 feet," "350–400 feet," and "400–450 feet." For it are recorded measurements all the way from 300 to 475 feet, a range of 175 feet when the distance to the next higher terrace is but 25 feet greater than the figure last mentioned. Terrace *e* (500 to 525 feet) is elsewhere repeatedly listed as "500–550 feet," and occasionally as "475–500" and "500–600," while actual figures referred to it range through 125 feet, when the distance to the next higher terrace is but 50 feet greater than the higher elevations assigned to terrace *e*. For terrace *f* (650–700 feet) a range is elsewhere given of "650–750 feet." Few elevations are given for terrace *g* (800–900 feet), but elsewhere this is more than once referred to as the terrace "at 800 feet," while gravel and sand at 784 feet are "accepted without any hesitation whatever" as a beach deposit and correlated with similar material found slightly above 800 feet. Little concrete evidence for terrace *h* with its estimated elevation of 1000–1200 feet is presented, although it is elsewhere referred to as a terrace "at 1000 feet."

It is evident from the wide range of altitudes and the conflicting elevations assigned to the several terraces that figures given in the above correlation table are highly generalized. From the text it is clear that to a considerable extent they represent estimated averages, made after allowing for an unknown depth of alluvial covering, an unknown amount of erosion or warping, or other uncertain factors. The author further makes clear that elevations greater than 50 feet are (except for occasional precise bench-mark altitudes) based on barometric observations and adds: "Since inaccuracies may arise with the most careful use of the barometer, and since it has generally been found difficult to determine exactly the position of the inner edge of the wave-levelled portion of terraces, the writer has often not defined their heights locally within 50 feet or so." Thus for three reasons the figures given for Coast *X* in the above correlation table are wholly unsuitable for correlation purposes: (1) They are based in considerable part on barometric observations with an estimated error of "50 feet or so," when the interterrace space often is but 100 feet, or even less; (2) they are inferred and highly generalized figures; (3) they cover ranges of altitudes so great, in comparison with the ranges of interterrace spaces, as to lose all value as proof of the assumed correspondence of terrace remnants. The figures cited for the Mediterranean Coast are also open to question, but those for Coast *X* are sufficient to illustrate the principles here discussed.

If, despite the three objections just enumerated (any one of which is seemingly fatal to trustworthy conclusions), one does seek to correlate the terrace-altitude figures from Coast *X* with those recorded from the Mediterranean Coast, due attention must be paid to the *relative* magnitude of discrepancies, rather than to the ab-

solute intervals separating one group of altitude figures from another. As is commonly the case in terrace-correlation studies, the paper in question appears to pay attention to the *absolute* number of feet separating one set of figures from its nearest neighbor in the other column. As the mean of one group in a majority of cases does not exceed the mean of the other by more than 26 feet, is in one case as little as 5 feet, and never exceeds 50 feet, while the nearest figure in one group never falls outside the figures of the other group by more than 34 feet and in half the cases is actually included within those figures, the correlator of these terraces could truthfully say: "There is . . . a fairly close accordance in the heights of the terraces representing each stage." But this is true only when consideration is given solely to the *absolute* distances separating the corresponding groups of altitude figures.

The *relative* positions of the altitude groups can be brought out effectively only by plotting the ranges to scale. This has been done in Figure 1. It there appears that 120–150 feet (terrace *b*) is relatively very far from 92–98 feet (terrace *b'*), when terraces are spaced at intervals of less than 100 feet; that under such circumstances 230–250 feet (terrace *c*) is anything but close to 181–198 feet (terrace *c'*), while 330–380 feet (terrace *d*) is markedly discordant with 295–328 feet (terrace *d'*). Instead of one of these three sets of terrace remnants according in elevation with the other, the members of each fall at or near the middle of the gaps between terraces of the other set. Instead of "a fairly close accordance" we have here *almost the maximum possible discordance.* If either of the sets of figures for terraces *bb'* and *cc'* ranged appreciably farther up or down in either column, the only result would be to bring them more closely into harmony with the next higher or next lower set of figures, with the result that an investigator would correlate the terraces entirely differently, but with even greater confidence than that with which the original correlation was made. The ranges for terraces *g* and *h* are so enormous, being greater than the interterrace spaces, that no significance can attach to the fact that terraces *g'* and *h'* fall somewhere within those generalized ranges. They might be close to or far from the real positions of terraces *g'* and *h'*. There remain terraces *a* and *a'* and terraces *e* and *e'* which are mildly discordant, and terraces *f* and *f'* which may agree or may be markedly discordant, according to the real position of terrace *f* within its rather large assumed range (50 feet in the tabulation, but 100 feet (650–750 feet) in the text). Thus an apparently strong case for correlation is seen, on careful analysis, to be very weak.

It should be obvious that if terraces falling halfway between other terraces can be correlated with those other terraces on the basis of altitude, then we must admit that *there can be no set of terraces in the world which cannot be correlated with any other set in the world having approximately the same range above sea level for the highest and lowest members of the series.* For halfway between terraces is the maximum possible discordance. All other altitude figures will fall still closer to the next higher or next lower terrace and, comparative altitude being the sole basis of correlation, must be correlated with the one or the other as the case may be. It need hardly be added that, when terraces at all possible altitudes can be correlated, the method of correlation is radically defective and can prove nothing.

It may be asked whether, in the columns of altitudes compared above, *the agree-*

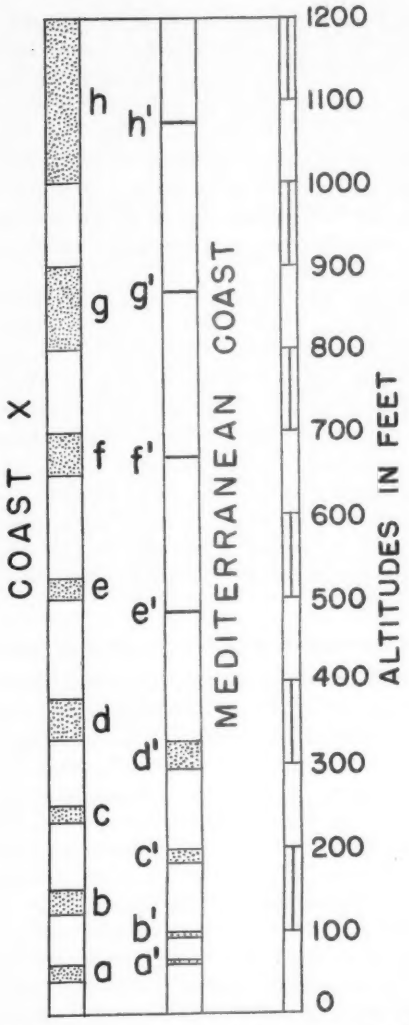


FIGURE 1.—*Terraces of Coast X and Mediterranean Coast plotted to scale*

ment in the number of altitude groups, if supplemented by some measure of agreement in magnitude of the vertical spaces between successive altitude groups, is not a sufficient basis for safe correlation of the terraces of Coast *X* with the Mediterranean terraces. The answer is that if altitudes have been recorded for truly significant elements of the terrace remnants; if the groups of altitude figures represent actual field measurements and not inferred, generalized, or standardized figures; if the ranges of altitude

figures grouped together are sufficiently restricted; if the total number of the restricted zones of altitude figures is actually identical in both regions (not the number selected for comparison, but the total number existing in the field); if each altitude group consists of *many* records of widely separated terrace remnants, none of them representing only one or a few such remnants; if the large intervals between the restricted zones referred to above are relatively or completely free from terrace remnants, so that there are really few or no altitudes to be reported from those intervals; and if the spacing between the altitude groups is reasonably identical in both cases; then the correlation rests on a foundation sufficiently firm to make it worthy of very serious consideration.

If the object of this paper were to prove or disprove the validity of a certain correlation between the terraces of Coast *X* and the Mediterranean terraces, it would be proper to enquire how far the altitude figures in the two columns earlier cited meet the requirements just stated. Our present purpose is very different: It is to discuss general principles of terrace correlation, citing actual instances only when necessary to illustrate and clarify those principles. But the reader, when examining any study of terrace correlation, whether marine or fluvial, should determine just what is back of all the figures employed and decide for himself whether they afford a safe basis for the correlations reported. In many cases where the writer has attempted such an analysis it has seemed to him that the sets of figures employed fail to meet the requirements outlined above as essential to safe correlation *where altitude is the sole basis of correlation.*

ESSENTIAL REQUIREMENTS OF CORRELATION BY ALTITUDE

Correlation of terraces, to be trustworthy, demands accurate determination of the altitudes of significant elements of terrace form or structure, or both; and the plotting (to some carefully selected vertical and horizontal scale) of the original field determinations of altitude, unaltered. Where the altitudes determined are known to be close to, but somewhat above or below, the true position of the significant terrace element, one should leave the original measurements on the graph but in addition plot, in some easily distinguishable form, the minimum and maximum possible altitudes of the significant feature, as judged by the observer, independently for each point, *while making the field examination.* This last is important not merely to get a judgment critically checked while the field evidence is still at hand, but also to avoid the inevitable tendency of the mind to correct figures (known to require correction) upward or downward in such manner as to bring them into harmony with what the observer believes other evidence shows to be the correct position. The procedure indicated will keep facts, and inferences based on facts, separate and distinct. This is essential to any sound scientific work, but it is doubly important in terrace-correlation studies.

Obviously there should be introduced into the picture only those inferences which have carefully been scrutinized and tested as to their soundness. If, for example, the alluvium on a wave-cut platform is of unknown thickness, or an exposed portion of the platform is an unknown distance out from the former marine cliff or a known distance so great that its position below the former sea level cannot be estimated with

any degree of accuracy, or the platform has been reduced an unknown amount by subsequent stream erosion, it is a serious error to determine an altitude at such a point, make a rough estimate or guess as to the probable former position of the platform, and plot resulting figures on the graph. Let the observer keep clearly in mind what he is trying to do. He is endeavoring to *prove, by altitude determinations alone,* whether certain terrace remnants do or do not belong together. Rough estimates and guesses have no probative value. They may easily betray the observer into a wrong correlation. They cannot possibly demonstrate the validity of a correct correlation.

If the precautions indicated are carefully observed, then the facts as to the precise position of the significant elements of certain terrace remnants, and the sound inferences as to the upper and lower limits between which the significant elements of other terrace remnants must fall, will be clearly before the investigator. The graph will tell its own story. If the evidence of observed facts alone is sufficient to demonstrate the existence of one, two, or more distinct and continuous terraces, the result of the study is conclusive. If observed facts plus apparently sound inferences indicate the existence of such terraces, the result obtained has a very high degree of probability. Little argument or explanation will be required. If the data on the graph are so scattered that argument and explanation are required to show that this figure is probably too high, that figure too low, and that certain possible and reasonable interpretations will bring the figures more nearly into harmony, the case for correlation is lost. The sole *proof* that the terrace remnants really correspond—identity of altitudes—is lacking. Under the conditions stated, there is not even that close similarity of altitudes which would give to any correlation a high degree of probability.

The precautions necessary when one is basing terrace correlation on identity of altitude are equally necessary when correlation is based on regularity of vertical spacing of terrace remnants. This latter method has been described elsewhere[3] and need not be discussed here.

DRAWING CONCLUSIONS FROM TERRACE CORRELATION

When the investigator of coastal terraces (1) has successfully determined, beyond reasonable doubt, that the forms he is attempting to correlate are of marine origin and that they fix with some degree of precision the positions of former sea levels, (2) has by careful instrumental leveling determined a sufficient number of precise or reasonably exact altitudes of parts of terrace remnants which have a known relation to such former sea levels, and (3) has determined by proper methods of correlating these altitude figures that there is valid proof of a certain specified number of distinct terraces, each representing a significantly long stand of the sea, he then has the equipment of observed facts which is the essential basis for a scientific analysis of the problem of sea-level changes.

The general principles involved in any such analysis, and the precautions which are necessary to insure valid conclusions, have been presented fully elsewhere.[4]

[3] Johnson, Douglas (1933) *The correlation of ancient marine levels,* Cong. Inter. Géog., Paris, 1931, C. R., Tome II, p. 42–54.

[4] Johnson, Douglas (1933) *The rôle of analysis in scientific investigation,* Geol. Soc. Am., Bull., vol. 44, 461–494.

Here we may point out that several hypotheses were long ago invented to account for facts of the type now under discussion. Differential uplift of the land due to orogenic movements, differential uplift of the land due to epeirogenic movements, eustatic drop of sea level due to withdrawal of ocean water forming ice caps and to other causes, or combinations of two or more of the above: these are among the multiple hypotheses which the investigator may adopt and test, and to which he is at liberty to add, if he can, other hypotheses invented by himself.

It may also be pointed out that deduction of the inevitable consequences of each hypothesis, and confrontation of those deduced consequences with observed facts in an effort to test the validity of the several hypotheses, has generally not been carried sufficiently far in papers dealing with coastal-terrace correlation. Too often the investigator is content, for example, to note that his figures resemble more or less closely those found on certain other coasts by other investigators and to pass at once to the conclusion that a eustatic drop of sea level explains the phenomena. This is a simple disposition of the matter. But to predict a world-wide history to explain the features of some coasts when evidence from other coasts may give no indication of such a history, or may even seem incompatible with such history, would not be to advance us toward the true solution of the problem. That there have been eustatic risings and fallings of sea level seems inevitably to follow from certain events in well-established earth history. But this should not lead us to over-hasty acceptance of the hypothesis that particular coastal terraces, formed at a particular time, were left far above sea level by that particular process. Uncritical acceptance of the hypothesis, and the pyramiding of other hypotheses and interpretations upon it, will only lead to confusion, if later discoveries prove that the known geological history of the same or other coasts excludes a drop of sea level at that particular time or indicates a wholly different cause for the present position of the ancient strands.

Attention may be directed to other possibilities of error incurred when the investigator concludes that the presence of certain more or less nearly horizontal terraces along an extended stretch of coast must be due to a eustatic drop of sea level. One frequently encounters in the literature such statements as the following: "Since a land area so large as France could not be raised uniformly throughout, the sea level must have dropped"; or "Uniform uplift of so large a group of islands in the Pacific being out of the question, we must conclude that sea level has fallen." Such statements seem open to criticism on two grounds. In the first place they usually, if not always, overstate the degree of approximation toward horizontality of the terraces which has definitely been proved by the evidence presented. Seldom, if ever, is the evidence of such character that absolute horizontality of the correlated terrace remnants can be demonstrated. This is important, for it is much easier to understand how widespread land movement could raise an extended coast more or less uniformly throughout than it is to believe that so extended an uplift could be absolutely uniform.

The second objection to such statements as those above quoted is that they unduly limit Nature. No one knows how large a portion of the earth's crust can be raised by approximately the same amount. Our acquaintance with the character, causes and effects of epeirogenic movements is not yet so perfect that we can state what, limits Nature sets to the area that can experience fairly uniform uplift. So, when

an investigator arbitrarily sets his own limits to what Nature can and cannot do in the respect, he may prematurely be closing the door on some hypothesis which might lead to the true explanation of the phenomena observed. Obviously the safer course is to keep the mind open and hospitable toward every hypothesis not definitely excluded by incontrovertible facts or by carefully scrutinized and fully tested deductions from such facts.

The student of river-terrace correlation must be equally careful in drawing conclusions from observed facts. He must, in the first place, be very sure of his facts and not confuse approximate uniformity in altitude of a terrace above present stream level with real uniformity. He must recognize that sloping terraces or benches representing remnants of former "valley-in-valley" forms do not fix former stream levels with any degree of accuracy and, in formulating hypotheses, he must make due allowance for all permissible former stream altitudes. He must not assume that a eustatic drop of sea level will cause streams everywhere to entrench a uniform distance below their former valley floors. Consideration of the dynamics of stream incision should convince him that the probabilities are otherwise. For he will realize that just as aggradation of a stream's course by exactly the same amount from its mouth to its headmost ravines is precluded by absence of any source of adequate debris in the latter areas, so uniform degradation is precluded by the special conditions under which incision of headwater ravines must take place. He will realize, further, that a stream's profile of equilibrium is a dynamic rather than a static thing; that the so-called "graded stream" is in reality a very slowly aggrading or very slowly degrading stream (normally the latter); that the gradient of the profile of equilibrium varies with every variation in climate, depth of incision, slope of valley walls, character of rock encountered, and other physical factors, and hence is forever changing; and that under these conditions it is in the highest degree improbable that at the present time such a constantly changing profile should happen to be, over any great stretch of territory, everywhere the same distance below another position of that changing profile as it chanced to be at the moment of incision.

If facts, determined with precision and established by incontrovertible evidence, demonstrate that a series of river terraces are approximately the same elevation above present stream level over considerable distances, or slope down valley more steeply or less steeply than does the present stream, conclusions based on those facts must be as cautiously drawn as in the case of coastal terraces. Some of the principles discussed above apply equally to stream terraces, such as, for example, the importance of not generalizing or overstating the degree of approximation toward horizontality of river terraces; the importance of not overstating the degree of approximation toward uniformity in the altitudes of river terraces above present stream level; the danger of drawing unwarranted conclusions from either of the above types of overstatement; and the danger of limiting Nature by assuming that uniformity of terrace altitudes above present stream levels over large areas necessarily implies a drop of sea level rather than an uplift of the land.

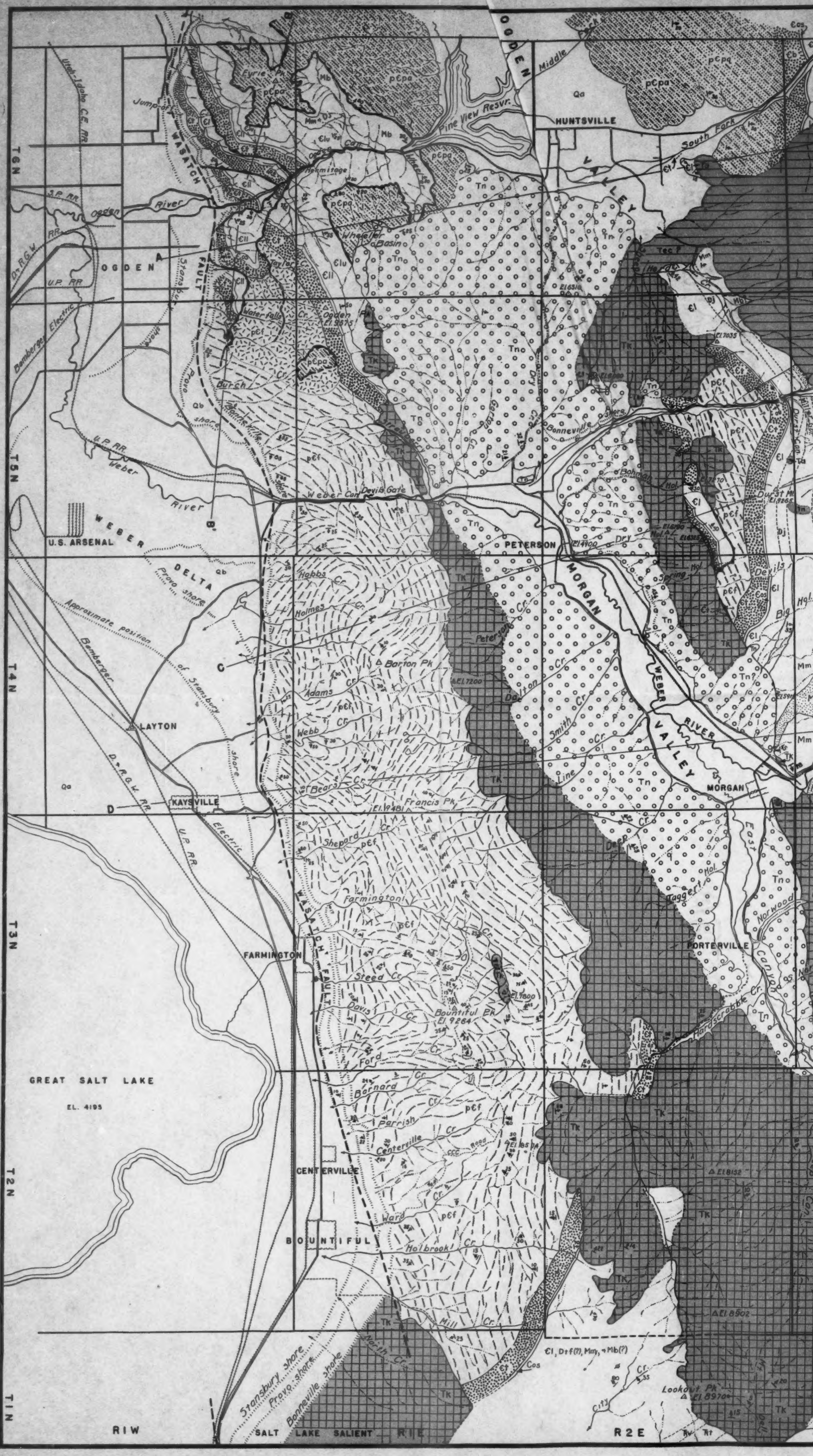
Other principles also enter into play when river terraces are involved. One cannot safely assume that river terraces above many or all streams in a region indicate either a change in land level or a change in sea level. The same phenomenon may

be produced by a change in climate, or it may represent nothing more than the normal history of streams lowering their channels as the supply of debris shed into them from land areas being gradually lowered is progressively decreased. Such phenomena may have wide regional development, and, if a chance approximation toward equality of elevation of some of the terraces above present stream level be overstated, an apparently sound basis for a false deduction of lowered sea level may be the result.

It must be further be remembered that streams cut slowly through resistant rock barriers while widening their valleys rapidly in weaker rocks, and that the contrast is often very remarkable. If, as commonly happens, a waterfall develops on the downstream side of a highly resistant barrier, the valley above the barrier may reach maturity in weak rock before the waterfall has retreated through the obstruction. When the fall finally cuts through the barrier and is rapidly retrograded in the weaker rock up stream, a young gorge in a mature valley, or a valley-in-valley form will result. This result is indistinguishable from individual valley-in-valley forms due to successive uplifts of the land or successive drops of sea level. The presence or absence of a barrier is not distinguishing, first because remnants of such forms may be found very remote from the barrier which caused them; and second because variations in rock resistance commonly exist in areas affected by uplifts of the land or lowering of sea level. Location of former stream levels in valley-in-valley forms cannot be effected with accuracy, and reconstruction of the ancient stream profiles involves a large measure of personal judgment. A fortuitous agreement in some restored levels, supplemented by *permissible* but not *probative* restorations in other areas, may give an apparent weight of evidence that seems conclusive when it is far from being so.

SUMMARY

No attempt has been made in these pages to treat systematically the whole question of terrace correlation. Effort has been rather to show that terrace correlation is a more difficult and delicate procedure than is generally realized. The writer has sought to demonstrate that, in the apparently simple steps of recognizing terraces of marine or fluvial origin, determining their altitudes, correlating the altitude figures, and drawing conclusions therefrom, there are numerous pitfalls to be avoided. It has been shown that there are special reasons why investigators in the field of terrace correlation should take unusual precautions to insure the validity of their results. If the foregoing discussion of difficulties and the means of overcoming them contains suggestions of help to other workers in this fascinating field of geomorphic research, the objects of the present essay will have been accomplished.


OGDEN
HUNTSVILLE
Eyrie Pk
Jump-off
Utah-Idaho C.E. RR
WASATCH
Middle
Pine View Resvr.
South Fork
VALLEY
Hermitage
Wheeler Basin
Wheeler
Taylor
Ogden River
U.P. RR
D.R.G.W. RR
Bamberger Electric
Stanley
Ogden El. 9675
Water Falls
Burch
Provo shore
Stansbury shore
Bonneville
Dry Canyon
Weber River
Weber Canyon
Devils Gate
U.S. ARSENAL
WEBER
DELTA
PETERSON
El. 4100
MORGAN
Hobbs
Holmes
Peterson
Barton Pk
El. 7200
Adams
Dalton
Webb
Bear
Francis Pk
El. 9481
Smith
Line
LAYTON
KAYSVILLE
Approximate position of Stansbury shore
Shepard
Farmington
FARMINGTON
Steed
Davis
Bountiful Pk
El. 9264
Ford
Bernard
Parrish
Centerville
CENTERVILLE
Ward
Holbrook
BOUNTIFUL
Mill
North
Deep
Taggert
East
Norwood
PORTERVILLE
Canyon
Hardscrabble Cr.
El. 7800
El. 8132
El. 8502
Lookout Pk
El. 8970
GREAT SALT LAKE
EL. 4195
SALT LAKE SALIENT
Stansbury shore
Provo shore
Bonneville shore
T5N
T4N
T3N
T2N
T1N
R1W
R1E
R2E

EXPLANATION

Qa,Qb — VALLEY ALLUVIUM, LAKE BONNEVILLE SEDIMENTS

ANGULAR UNCONFORMITY

Tn — NORWOOD TUFF

ANGULAR UNCONFORMITY

WASATCH GROUP

Tk — KNIGHT (?) FOR.

CONFORMABLE (?)

Tf — FOWKES (?) FOR.

CONFORMABLE (?)

Tsm — SAW MILL CONGL.

ANGULAR UNCONFORMITY

Tpc — PULPIT CONGL.

Ta — ALMY CONGLOMERATE EQUIVALENT TO BOTH Tpc AND Tsm

ANGULAR UNCONFORMITY

KTh — HENEFER FOMATION

ANGULAR UNCONFORMITY

Kf / Hol — FRONTIER AND ASPEN FORMATIONS SANDSTONE RIDGE-MAKERS STIPPLED

DISCONFORMITY

KK — KELVIN (?) FOR.

DISCONFORMITY

Je / Jtc — ENTRADA (?) SS. TWIN CREEK FOR.

DISCONFORMITY (?)

Jn — NUGGET SANDSTONE

DISCONFORMITY (?)

Ta — ANKAREH FORMATION

Trt — THAYNES LIME

Trw — WOODSIDE SH

Ppc — PARK CITY FO

Pw — WEBER QUART

Pm — MORGAN SAND

Mb — BRAZER FORMA

DISCONFORMITY (?)

Mm — MADISON LIMES

DISCONFORMITY (?)

Dtf — THREE FORKS (

DISCONFORMITY

€lu / €l — CAMBRIAN LMS. €lu, UNITS 3– €l, UNITS 1

DISCONFORMITY

Eos — OPHIR SHALE

DISCONFORMITY

Ct — TINTIC QUARTZI

DISCONFORMITY (?)

p€p — PROTEROZOIC (?) S p€pc – LOWER AR GEOUS STRA p€pq – UPPER QU STRATA

ANGULAR UNCONFORMITY

p€f — FARMINGTON CAN COMPLEX

WAVY LINES SHOW FICATION WHERE K

THRUST FAULT. P HACHURES DIRECT UNDER THRUST

U/D — SMALL HIGH-ANGL FAULTS

U/D — BASIN AND RANGE FAULTS

BEDS HORIZONTAL

BEDS VERTICAL

LAKE BONNEVILLE BEACH

LINE OF OUTCROP SOME PROMINENT AS TRACED ON AE PHOTOGRAPHS

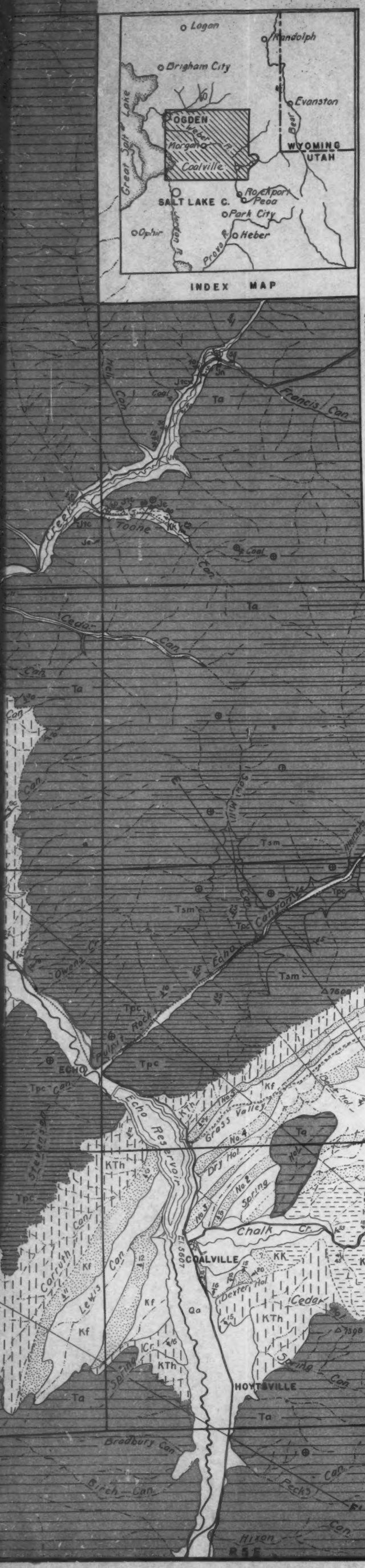
BASE COMPILED FROM AERIA PHOTOGRAPHS BY THE SOIL CO VATION SERVICE, THE FOREST S AND THE A.A.A.; ALSO FROM TH COALVILLE AND FORT DOUGLAS RANGLES, THE U.S. RECLAMATION VICE TOPOGRAPHIC MAPS OF WE COUNTY, THE COUNTY ENGINEER' MAP OF MORGAN COUNTY, U.S. EST SERVICE MAPS, AND PLAN TABLE MAPS OF COMPLICATED

BY A. J. EARDLEY

SCALE IN MILES

0 1 2 3 4 5

GEOLOGIC MAP AND CROSS SECTIONS OF THE NORTH-CENTRAL WASATCH MOUNTAINS, UTAH



EXPLANATION

Qa,Qb	VALLEY ALLUVIUM, LAKE BONNEVILLE SEDIMENTS	Tht	THAYNES LIMESTONE
ANGULAR UNCONFORMITY			
Tn	NORWOOD TUFF	Tw	WOODSIDE SHALE
ANGULAR UNCONFORMITY			
Tk	KNIGHT (?) FOR.	Ppc	PARK CITY FORMATION
CONFORMABLE (?)			
Tf	FOWKES (?) FOR.	Pw	WEBER QUARTZITE
CONFORMABLE (?)			
Tsm	SAW MILL CONGL.	Pm	MORGAN SANDSTONE
ANGULAR UNCONFORMITY		*DISCONFORMITY*	
Tpc	PULPIT CONGL.	Mb	BRAZER FORMATION
		DISCONFORMITY (?)	
Ta	ALMY CONGLOMERATE EQUIVALENT TO BOTH Tpc AND Tsm	Mm	MADISON LIMESTONE
ANGULAR UNCONFORMITY		*DISCONFORMITY (?)*	
KTh	HENEFER FOMATION	Dtf	THREE FORKS (?) FOR.
ANGULAR UNCONFORMITY		*DISCONFORMITY*	
Kf	FRONTIER AND ASPEN FORMATIONS SANDSTONE RIDGE-MAKERS STIPPLED	€lu / €ll	CAMBRIAN LMS. & DOL. €lu, UNITS 3 - 13 €ll, UNITS 1 & 2
DISCONFORMITY			
Kk	KELVIN (?) FOR.	€os	OPHIR SHALE
DISCONFORMITY			
Je / Jtc	ENTRADA (?) SS. TWIN CREEK FOR.	€q	TINTIC QUARTZITE
DISCONFORMITY (?)			
Jn	NUGGET SANDSTONE	pCp	PROTEROZOIC (?) STRATA pCpa - LOWER ARGILLACEOUS STRATA pCpq - UPPER QUARTZITIC STRATA
DISCONFORMITY (?)		*ANGULAR UNCONFORMITY*	
Ta	ANKAREH FORMATION	pCf	FARMINGTON CANYON COMPLEX WAVY LINES SHOW STRATIFICATION WHERE KNOWN

WASATCH GROUP

THRUST FAULT. POINTED HACHURES DIRECTED UNDER THRUST SHEET

U/D — SMALL HIGH-ANGLE FAULTS

U D — BASIN AND RANGE FAULTS

⊕ BEDS HORIZONTAL

⊖ BEDS VERTICAL

LAKE BONNEVILLE BEACH

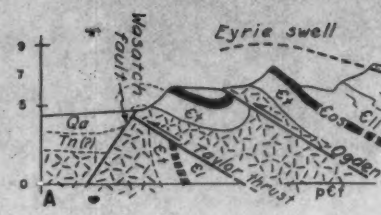
LINE OF OUTCROP OF SOME PROMINENT BED AS TRACED ON AERIAL PHOTOGRAPHS

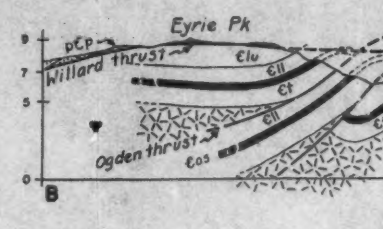
BASE COMPILED FROM AERIAL PHOTOGRAPHS BY THE SOIL CONSERVATION SERVICE, THE FOREST SERVICE, AND THE A.A.A.; ALSO FROM THE COALVILLE AND FORT DOUGLAS QUADRANGLES, THE U.S. RECLAMATION SERVICE TOPOGRAPHIC MAPS OF WEBER COUNTY, THE COUNTY ENGINEER'S MAP OF MORGAN COUNTY, U.S. FOREST SERVICE MAPS, AND PLANE TABLE MAPS OF COMPLICATED AREAS.

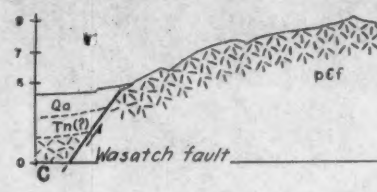
BY A. J. EARDLEY

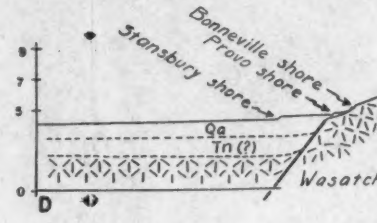
SCALE IN MILES

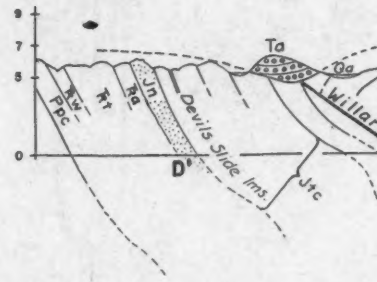
0 1 2 3 4 5

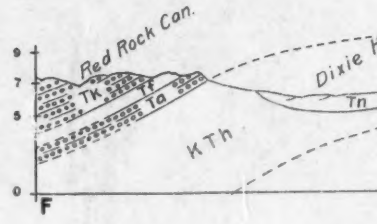












Ogden thrus

Taylor thrus

IC MAP AND CROSS SECTIONS OF THE NORTH-CENTRAL WASATCH MOUNTAINS, UTAH

swell
pCp
Mb
Huntsville sag
Tn
Qa
Ta (?)
Brigham qtz
Ct
Cos
Cll
Clu
Dtf(?) + Mm(?)
Willard thrust
Durst thrust(?)
Sea level
Ogden thrust
pCf
A'

Ogden thrust
Willard thrust
Taylor thrust
Ct
Qa
Tn (?)
pCf
Sea level
B'

Durst Mt.
Cl
Cos?
Dtf(?)
Mm
Mb
Pu
pCp(?)
Willard thrust
Morgan Valley
Qa
Tn
Tk
Durst thrust
Morgan block
Sea level
C'

Pediment
Pediment
Morgan Valley
App. Tk contact to south
Wasatch fault
Sea level
Pw
Ppc
Pm
Mb
Mm
Dtf
Rw
Rt
Ra
Jn
Jtc
D'

Tsm
Tpc
Ta
KTh
Kf
Echo Can.
Sea level
E
E'

Qa
Ta
Echo Can.
Willard thrust
Unconformity
KTh
Kf
KK(?)
Sea level
D"

Dixie Hol.
Tn
KTh
Kf
Qa
Ta
Weber Valley
Sea level
F'

VERTICAL AND HORIZONTAL SCALE IN MILES
0 1 2 3 4 5

FORMATIONAL DESIGNATIONS

- Qa VALLEY ALLUVIUM
- Tn NORWOOD TUFF
- Tk KNIGHT (?) FORMATION
- Tf FOWKES (?) FORMATION
- Tsm SAW MILL CONGLOMERATE
- Tpc PULPIT CONGLOMERATE
- Ta ALMY CONGLOMERATE
- KTh HENEFER FORMATION
- Kf FRONTIER FORMATION
- Ka ASPEN FORMATION
- Kk KELVIN FORMATION
- Jtc TWIN CREEK FORMATION
- Jn NUGGET SANDSTONE
- Ra ANKAREH FORMATION
- Rt THAYNES LIMESTONE
- Rw WOODSIDE SHALE
- Ppc PARK CITY FORMATION
- Pw WEBER QUARTZITE
- Pm MORGAN SANDSTONE
- Mb BRAZER FORMATION
- Mm MADISON FORMATION
- Dtf THREE FORKS (?) FORMATION
- Clu UPPER CAMBRIAN LIMESTONE
- Cll LOWER CAMBRIAN LIMESTONE
- Cl CAMBRIAN LIMESTONE,
- Cos OPHIR SHALE
- Ct TINTIC QUARTZITE
- pCp PROTEROZOIC (?) STRATA
- pCf FARMINGTON CANYON COMPLEX

Eyrie swell
Willard thrust
Huntsville sag
Qa + Tn
pCp
Cos
Cl
Dtf(?)
Ct
pCf

SECTION ALONG THE NORTH SIDE OF OGDEN CANYON AND ACROSS OGDEN VALLEY

Huntsville sag
Willard thrust surface
ENE
Sea level
Ogden thrust surface
Taylor thrust surface
Ogden Can.
Taylor Can.
NNW

BLOCK DIAGRAM SHOWING THE THREE THRUST SURFACES OF OGDEN CANYON AREA. THE FRONT SIDE IS SECTION B–B'

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 55, PP. 819-894, 14 PLS., 10 FIGS. JULY 1944

GEOLOGY OF THE NORTH-CENTRAL WASATCH MOUNTAINS, UTAH

BY A. J. EARDLEY

CONTENTS

	Page
Abstract	821
Introduction	821
Location of area and purpose of study	821
Acknowledgments	823
Stratigraphy	823
Pre-Cambrian rocks	823
Farmington Canyon complex	823
Proterozoic (?) rocks	827
Cambrian system	827
Tintic quartzite	827
Ophir shale	828
Cambrian limestone and dolomite	828
Devonian (?) system	830
Three Forks (?) formation	830
Mississippian system	831
Madison limestone	831
Brazer formation	831
Pennsylvanian and Permian systems	832
Morgan sandstone	832
Weber quartzite	833
Park City formation	834
Triassic system	835
Woodside shale	835
Thaynes formation	835
Ankareh formation	835
Jurassic system	835
Nugget sandstone	835
Twin Creek formation	835
Entrada (?) and Curtis (?) formations	837
Cretaceous system	838
Kelvin (?) formation	838
Aspen and Frontier formations	839
Montana or Paleocene series	840
Henefer formation (new name)	840
Paleocene series	842
Almy conglomerate	842
Fowkes (?) formation	844
Eocene series	844
Knight (?) formation	844
Oligocene series	845
Norwood tuff (new name)	845
Pleistocene and Recent deposits	847
Lake Bonneville sediments	847
Valley alluvium	847
Structure	847
General features	847

Thrusts 847
Ogden Canyon thrust area 847
Ben Lomond thrust area 850
Durst Mountain thrust area 852
Evolution of thrust structure 853
Folds 854
Coalville anticline 854
Stevenson Canyon syncline 855
Henefer anticline 855
Morgan Valley syncline 856
Folds of the central Wasatch 856
High-angle faults of Laramide age 857
Superposition of structures 858
Sequence of diastrophic stages 859
Relation of local structures to regional 863
Time relations 863
Special relations 866
Physiography 872
Late Laramide surfaces 872
Herd Mountain surface 873
Weber Valley surface 874
Age of the erosion surfaces 877
Basin and Range faulting 880
Wasatch front 880
Salt Lake salient 881
Pleasant View salient 883
Dissection of the Weber Valley surface 886
Gilbert's theory 886
Lake Bonneville terraces and deposits 887
The Wasatch front 887
Morgan Valley and the Weber delta 888
Ogden Valley and the artesian reservoir 888
Prospects of ground water in Morgan Valley 889
Ground water in the Great Salt Lake Basin 889
Post-Bonneville events 890
Conclusions 891
References cited 892

ILLUSTRATIONS

Figure — Page
1. Index map of the north-central Wasatch Mountains 822
2. Correlation diagram of diastrophic episodes 864
3. Fault map of north-central Utah 866
4. Hypotheses of thrust structure 868
5. Herd Mountain and Weber Valley surfaces 875
6. Origin of the Herd Mountain surface 876
7. Profiles of the west front of the Wasatch Mountains 877
8. Origin of the Salt Lake salient 882
9. Development of a re-entrant in a fault scarp 884
10. Origin of the Pleasant View salient 885

Plate — Facing page
1. Geologic map and cross sections of the north-central Wasatch Mountains, Utah 819
2. Geologic map and cross sections of the Ben Lomond area, Wasatch Mountains, Utah 850
3. Tectonic map of the north-central Wasatch 858

4. Aerial mosaic of the Ogden Canyon area 874
5. Shapes and relations of the thrust surfaces 875
6. Panoramas of Ogden Canyon and Morgan Valley 876
7. Relief models of the north-central Wasatch 877
8. Salt Lake salient 878
9. High Herd Mountain surface 879
10. Pediment in Lost Creek Valley 882
11. Pediment in Weber Valley 883
12. Weber Valley surface on Antelope Island 886
13. Dissection of Weber Valley surface 887
14. Lake Bonneville deposits 890

ABSTRACT

The formations used in dating the stages of deformation of the north-central Wasatch are the Twin Creek, Entrada (?), and Curtis (?) (Jurassic); the Kelvin (?), Aspen, and Frontier (Cretaceous); the Henefer (new name: lower Paleocene or uppermost Montana); the Wasatch group consisting of the Pulpit conglomerate (new name), Saw Mill conglomerate (new name), Fowkes (?) formation, and Knight (?) formation (middle Paleocene to middle Eocene); and the Norwood tuff (new name: lower Oligocene). Stages of deformation recognized are: (1) uplift and perhaps thrusting in a western highland at the beginning of Kelvin time, (2) thrusting in post-Frontier—pre-Henefer time (Taylor thrust), (3) renewed thrusting in pre-Henefer time (Ogden thrust), (4) crossfolding of thrust sheets and first major elevation of Uintas, also in pre-Henefer time, (5) renewed thrusting in post-Henefer—pre-Pulpit time (Willard thrust), (6) gentle crossfolding of Willard thrust sheet and folding of Pulpit conglomerate around Coalville anticline in pre-Saw Mill time, (7) renewed gentle folding prior to deposition of Norwood tuff, (8) gentle folding in post-lower Oligocene, erosion of Herd Mountain surface during the Miocene, (9) uplift, rejuvenation, and erosion of Weber Valley surface, a pediment, (10) Basin and Range faulting followed by dissection and alluviation of the pediment, the rise and fall of Lake Bonneville, and post-Bonneville dissection.

The structures trend in three directions, and problems concerning their superposition are discussed. The local structures are correlated with the regional from the standpoint of chronological development and of distribution and orientation. Tuffaceous beds indicate volcanic activity at five times in adjacent areas.

INTRODUCTION

LOCATION OF AREA AND PURPOSE OF STUDY

The Wasatch Mountains extend from central Utah northward to the Idaho boundary. Salt Lake City is at their west base and about midway of their length. The portion designated the north-central Wasatch in this report embraces an area from Bountiful 40 miles northward to North Ogden, and from Great Salt Lake 36 miles eastward to Coalville. In terms of topographic landmarks the area extends from the Salt Lake salient[1] on the south to the Pleasant View salient on the north, and from the Great Salt Lake Valley on the west to the mouths of the Echo and Chalk Creek canyons, tributaries of the Weber River Valley, on the east. The extent of the area studied is shown on Figure 1 and on the index map of Plate 1. The Weber River runs from the southeast corner to the northwest corner, and most of the area is drained by it. In its lower reaches it passes through a V-shaped gorge called Weber Canyon about 4000 feet deep before reaching the Great Salt Lake Valley. Another gorge between Morgan and Croyden is designated the upper Weber Canyon. The wide part of the Weber Valley between the two canyons will be

[1] The mass of hills that projects westward from the Wasatch front just north of Salt Lake City has been called the "City Creek spur" by Gilbert (1928), and among other geologists and local residents the hills are spoken of as the City Creek salient, the Salt Lake salient, and Beck's spur. To geologists having only a casual acquaintance with the region the geographic term "Salt Lake salient" seems most informative and is, therefore, used in this report. See plates 1, 7.

referred to as Morgan Valley. A widening of upper Weber Canyon about 3 miles above Morgan is known as Round Valley. The Weber Valley above upper Weber Canyon is known as Henefer Valley from the town of Henefer. All these features are at once recognizable on the relief model, Plate 7. Four main tributaries enter

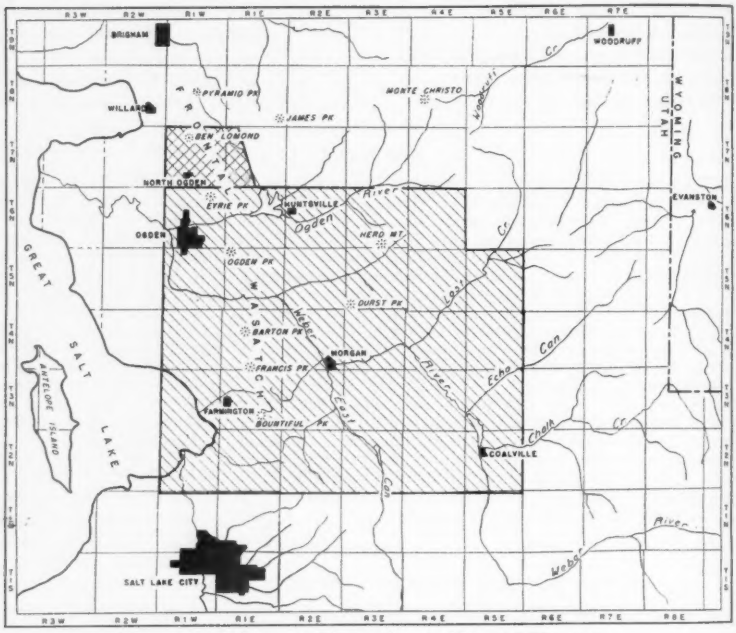


FIGURE 1.—*Index map of the north-central Wasatch Mountains*

Diagonally lined area is that of the geologic map, Pl. 1, and cross-ruled area is that of the geologic map, Pl. 2.

the Weber on the right going downstream. They are, from southeast to northwest, Chalk Creek, Echo Canyon Creek, Lost Creek, and Cottonwood Creek. East Canyon Creek enters on the left at Morgan. At the very north end of the area Ogden River flows through Ogden Canyon, a gorge similar to Weber Canyon. Above Ogden Canyon is the agricultural basin known as Ogden Valley into which several forks of Ogden River discharge. Ogden Valley and Morgan Valley are separated by a low divide and together form a large north-south depression second in size only to the Great Salt Lake Valley. Between these two great depressions stretches a mountain ridge which supports the highest peaks of the region. This high ridge is referred to as the frontal Wasatch to distinguish it from the rest of the north-central Wasatch area.

The primary purpose of the study has been to describe the structures of the area, to decipher the stages of their growth, to relate the local structures in time and space to the regional structure, and in so doing to gain some appreciation of the forces that

caused them. A secondary purpose has been to work out additional details of the erosional and depositional history in post-Laramide time. As the late erosional and depositional history is intimately connected with the Basin and Range faulting this aspect of the structure is treated under the heading of physiography.

Since the structural study is concerned chiefly with the Jurassic, Cretaceous, and Tertiary disturbances, the pre-Jurassic stratigraphy has only been dealt with sufficiently for structural purposes. Most of the field work centered about the younger formations, and a number of stratigraphic contributions are made regarding them.

ACKNOWLEDGMENTS

During the field work which progressed through short and long summer periods from 1933 to 1942, the writer was assisted by several advanced and graduate students at the time in the University of Michigan, including Max Demorest, R. A. Hatch, Kenneth McGlone, Robert Culbertson, and John Durr. R. A. Hatch of the Department of Mineralogy, University of Michigan, studied several thin sections; John Clark of the Carnegie Museum visited the writer for 2 days in the area, collected some Oligocene vertebrate fossils, and reported their identifications; Chester A. Arnold of the Museum of Paleontology of the University of Michigan collected and identified some plant fossils; J. Stewart Williams of the Utah State Agricultural College went over the Carboniferous and Devonian strata in Ogden Canyon with the author and assisted materially in deciding upon boundaries and nomenclature; Hyrum Schneider spent 2 days with the writer, one in Ogden Canyon on the lower Paleozoic rocks, and one in Weber Valley on the Jurassic rocks; and V. Brown Monnett on the basis of 6 weeks' work on the Jurassic and Cretaceous rocks in Weber Valley at Coalville and Peoa supplied the writer with certain appreciated although tentative data. M. V. Denny's aid in preparing the photographs for the plates has been of much value.

Partial financial assistance for the field work came from the Faculty Research Fund of the University of Michigan and the Penrose Bequest of The Geological Society of America.

STRATIGRAPHY

PRE-CAMBRIAN ROCKS

Farmington Canyon complex.—The oldest rock exposed in the area is a stratified crystalline complex of metamorphic sedimentary rocks, metamorphic silicic igneous rocks, injection gneisses, and metamorphic mafic rocks (Eardley and Hatch, 1940a, p. 58–72). The crystalline complex underlies another ancient sequence which is probably Upper Proterozoic and in part Lower Cambrian. Table 1 outlines these relations and those of the other formations.

The sedimentary rocks of the ancient complex are now metaquartzites, quartz schists, arkosites, conglomerate schist, and metamorphosed graywacke. After sediments whose thickness exceeded 10,000 feet were deposited, they were thoroughly injected by sills and dikes during a prolonged orogeny which resulted in widespread development of foliation and various cataclastic structures. The prevailing strike of the foliation is north to northwest and generally parallels bedding.

TABLE 1.—*Stratified rocks of the north-central Wasatch Mountains*

Age	*Formation*		*Thickness* (Feet)	*Lithology*
Pleistocene and Recent	Valley alluvium Lake Bonneville sediments		50–1000+	Delta, beach, and bottom deposits of Lake Bonneville; post-Bonneville valley alluvium, alluvial fans, and sediments in Great Salt Lake.
Lower Oligocene (Chadronian)*	Norwood tuff		50–2000	Light-colored tuff with lenses of volcanic conglomerate becoming numerous to south. Vertebrate fossils found in beds of quartz sand with white powder, numerous gray quartzite pebble lenses and poorly consolidated tuff.
Lower Eocene (Wasatchian (?))*	Wasatch group	Knight formation	5000±	Deep-red series of conglomerates, sandstones, and shales.
Upper Paleocene (?)* (Clarkforkian and Tiffanian)	Wasatch group	Fowkes formation	800	Light-gray grit, sandstone, and shale, in large part of volcanic origin. Little indurated. Some shales black and paper thin. Fossil plants in layer 300 feet from bottom.
Middle Paleocene (?)* (Dragonian and Torrejonian)	Wasatch group	Almy conglomerate: Saw Mill conglomerate	1200	Lower 400 feet of massive, red, cliff-making conglomerates. Middle 500 feet of light-tan and gray conglomerates, grits, sandstones, and shales. Upper 300 feet of brownish-tan conglomerates.
	Wasatch group	Almy conglomerate: Pulpit conglomerate	2000±	Coarse, red, cliff-making conglomerate. Boulders are chiefly Cambrian and Pennsylvanian quartzites. Same is true for overlying Saw Mill conglomerate.
Lower Paleocene, post-Montana, or Upper Montana	Henefer formation		1250–10,000	Red, pink, purple, and light-gray shales. Red and gray sandstones with numerous lenses of pea and pebble conglomerate. Also gray, tan, brown, and red pebble and boulder conglomerate. 200-foot bed of volcanic tuff and lumpy limestone near bottom.
Upper Cretaceous (Colorado)	Frontier formation		10,000±	Coal-bearing series of hogback-making gray sandstones separated by dark shales and friable sandstones. A coarse conglomerate in middle of section.
	Aspen formation		250±	Black shale and light-tan sandstone. Some bentonite and procellanitelike beds present. Fish scales in black shale.

TABLE 1—*Continued*

Age	*Formation*	*Thickness* (Feet)	*Lithology*
Lower Cretaceous (?)	Kelvin (?) formation	3000±	An upper interbedded sequence of variegated shales and sandstones 2300 feet thick. A lower sequence containing four groups of conglomerate beds in light-gray sandstones and shales, 700 feet thick. Shales in lower unit are salmon, pinkish white and purplish red to brick red.
Upper Jurassic	Curtis (?) formation	Absent and up to 250	Gray to dark-gray shale and tan to greenish-tan sandstone. Speckled in part.
	Entrada (?) formation	1170	Deep dull-red shale, sillstone, and sandstone beds that weather into a soft red soil. Some sandstones speckled with gray grains.
	Twin Creek formation	3700±	Interbedded series of light greenish-gray and tan-gray limestones and shales. Shales are mostly calcareous, and some of limestones are argillaceous. Some limestones lithographic, some fossiliferous.
Middle or Lower Jurassic	Nugget sandstone	1500	Pink to salmon-colored sandstone, massively bedded and cliffmaker.
Upper Triassic (?)	Ankareh formation	1100	Mostly shale but with numerous sandstone beds up to 4 inches thick. Colors are dull red, dark purplish red, and gray-green. Ripple-marked throughout.
Middle Triassic	Thaynes limestone	2200	Dark-gray, medium-gray, and tan-gray limestones, fine-grained to crystalline. Drab-gray shale beds intercalated, and near the top is an olive and red splintery shale unit.
Lower Triassic	Woodside shale	600	Dark, brick-red, and sandy shale. Some shale beds are micaceous. A few thin sandstone beds are intercalated. Ripple marks common throughout.
Permian and Pennsylvanian (?)	Park City formation	1700	Light- to dark-gray limestones and sandy limestones, gray to olive to purplish shales, calcareous sandstones, great masses of dark chert, and phosphatic limestones and shales. Some of phosphatic limestones are oölitic.

TABLE 1—*Continued*

Age	*Formation*	*Thickness* (Feet)	*Lithology*
Pennsylvanian	Weber quartzite	3000±	Gray to white but buff-weathering quartzites and sandstones. Some gray and light-gray limestones carrying chert nodules present. Also a bed of chert near top of formation.
Pennsylvanian (?)	Morgan sandstone	1000	Pink to red sandstone that turns brick-red on weathered surface. Fairly friable. Some shale and a few thin limestone beds at top.
Upper (?) Mississippian	Brazer formation	1100	Limestones and sandstones, mostly gray and creamy gray. Lower 600-foot sandstone mostly tan and red. Purple and tan shale unit near top. Fossiliferous.
Lower (?) Mississippian	Madison limestone	650–800	Fossiliferous dark-gray limestone, generally thin-bedded, cherty toward top. Vertical jointing conspicuous. Cliff-maker.
Devonian (?)	Three Forks (?) formation	500–775	Shales and thin-bedded sandstones, red, tan, and drab. Some thin gray quartzitic sandstones and some pink to yellow sandy and shaly limestones present. Mostly a slopemaker.
Middle and Upper (?) Cambrian	Cambrian limestones and dolomites	1375±	Dominantly limestones but with shales and dolomites interbedded. Characteristic lithologies such as Bluebird type, Lynch type, and Hartmann type.
Middle Cambrian	Ophir shale	200	Brown to olive-green shale with micaceous sheen and large wormy markings on surfaces. Arkosic and quartzitic layers ¼ to 1 inch thick are abundant.
Lower (?) Cambrian	Tintic quartzite	500–1000	Dominantly pink and gray quartzite, but with some beds of buff, red, and blood-purple varieties. Intraformational conglomerate beds are common. A basal conglomerate 100 feet thick occurs in Cottonwood Canyon but not in Ogden Canyon.

TABLE 1—*Concluded*

Age	*Formation*	*Thickness* (Feet)	*Lithology*
Upper Proterozoic (?)	Proterozoic (?) rocks	10,000+	Interbedded series of phyllites, tillites arkosites, and quartzites. Upper 7000 feet dominantly purplish and rusty-weathering quartzites. Lower 3000 feet arkosites and phyllites interbedded in purple quartzites and metaquartzites.
Middle (?) pre-Cambrian	Farmington Canyon complex	10,000+	Stratified crystalline complex of metaquartzite, conglomerate schist, quartz schist, arkosite and graywacke, and injection gneiss.

* North American provincial ages from Wood *et al.* (1941).

An open anticline trends about N. 60° W. through Bountiful Peak (Pl. 1) and is flanked on both sides by open synclines, all of which pitch about 15°–20° NW. Elsewhere the structure is complicated and has not been satisfactorily solved.

Proterozoic (?) rocks.—A great series of interbedded quartzites, phyllites, tillites, and arkosites, much less metamorphosed than the underlying Farmington Canyon complex, has been described in the northern Wasatch (Eardley and Hatch, 1940b, p. 795–843), and the same sequence extends southward into the area of this report. In the mountain mass northeast of Ogden Valley, which is transected by the South and Middle forks of Ogden River and Wolf and Geertsen creeks (beyond the limits of the map), two divisions of Proterozoic beds were recognized. Approximately 7000 feet of strata, dominantly purplish and rusty-weathering quartzitic sandstones, overlies about 3000 feet of arkosites and phyllites interbedded with gray to green to purple quartzites. Part of the lower group is also exposed above the Willard thrust at the head of Ogden Canyon.

The sediments were probably deposited in a deep trough in Late Proterozoic and Early Cambrian times (Eardley and Hatch, 1940b, p. 829–831) and overlap the Archeozoic rocks of the northern Utah highland. They are probably separated from the overlying Cambrian strata by an unconformity, but in the trough area the unconformity was not found except in the Cottonwood uplift of the central Wasatch. The trough sequence was thrust over the thin near-shore Middle Cambrian deposits of the highland during the Laramide revolution. It is tentatively concluded that the metamorphism of the quartzites, phyllites, and arkosites was accomplished during the Laramide revolution, and not during the post-Proterozoic and pre-Middle Cambrian disturbance, because (1) the Laramide revolution was much more intense, and (2) burial at the time of the Laramide revolution was deeper.

CAMBRIAN SYSTEM

Tintic quartzite.—The Tintic quartzite is predominantly pink or gray, but buff, red, and dark-purple varieties occur. It is made up of medium-sized to coarse well-

rounded, glassy-appearing sand grains cemented by silica. The secondary silica is in optical continuity with the quartz grains, but in the pink and red quartzites a thin iron oxide border surrounds most of the grains and enables the observer to distinguish the original grains from the added material. Other minerals besides quartz constitute less than 1 per cent of the rock.

Intraformational conglomerates are common. The pebbles are either vein quartz or varieties of Proterozoic (?) quartzites that crop out in surrounding areas. A coarse basal conglomerate fully 100 feet thick shows a sedimentary contact with the Farmington Canyon complex in Cottonwood Canyon (north of Durst Mountain). The pebbles range from 1 to 3 inches in diameter, are well rounded, and consist of white, pink, and laminated gray quartzite (probably varieties of near-by Proterozoic (?) quartzite), vein quartz, and jasper. This basal conglomerate is thinner in Hardscrabble Canyon and absent in Ogden Canyon.

The Tintic quartzite is approximately 1000 feet thick in the Cottonwood-Durst Peak area, but only 500 to 700 feet in the front range of the Wasatch. It rests unconformably upon the crystalline complex. Its relation to the Proterozoic (?) strata is believed to be both one of conformity and disconformity (Eardley and Hatch, 1940, p. 835–837) from observations in surrounding areas, but no contacts are exposed in the area of this report.

The quartzite and the overlying Ophir shale and Cambrian limestone and dolomite are probably continuous from the Ophir district of the Oquirrh Mountains and the southern part of the Wasatch Mountains northward to this area. On the other hand, the Brigham quartzite whose type locality is in the Wasatch northeast of Brigham City is separated from the Ogden Canyon exposures of quartzite by several great thrusts and is much thicker. It is probably continuous with the exposures to the east in the Bear River Range. The term Tintic is, therefore, preferred for the north-central Wasatch exposures. Both the Tintic and Brigham quartzites are somewhat older than middle Middle Cambrian according to Deiss (1938, p. 1141, Fig. 5), and probably Lower Cambrian according to Williams and Maxey (1941, p. 276–277) and Wheeler (1943, p. 1811–1815).

Ophir shale.—The Ophir shale is brown to olive green with a characteristic micaceous sheen and large wormlike markings on bedding surfaces. Arkosic and quartzitic layers, $\frac{1}{4}$ to 1 inch thick, are common. Some paper-thin shales and oölitic limestone beds were seen near the mouth of Durst Canyon. On the high ridge extending westward from Durst Peak the shale is olive green, fairly thin-bedded, in part splintery, and only slightly micaceous.

No fossils were found in the area of this report, but the *Glossopleura* fauna occurs elsewhere in the lower part of the formation. The lithology is characteristic of the type locality, Ophir, Utah, and the stratigraphic position is the same. Correlation with the type locality therefore seems justified. According to J. Stewart Williams (personal communication) correlation with the Spence shale of northern Utah would also be acceptable. The formation is about 200 feet thick.

Cambrian limestone and dolomite.—A sequence of over 1300 feet of limestone and dolomite overlies the Ophir shale in the Durst Peak area and in lithology and stratigraphic sequence matches the beds overlying the Ophir shale in the Ophir district

fairly well. A section measured along the crest of the high ridge that extends westward from Durst Peak is given below:

Units		Thickness (Feet)
13	Limestone, dark gray at bottom becoming light gray at top. Vermicular (Bluebird type)	100±
12	Shale, tan thin-bedded	20
11	Dolomite, white on weathered surface, gray on fresh surface. Wavy fine laminations (Lynch dolomite type). Becomes darker and vermicular at top	50
10	Limestone, dark and pisolitic at bottom, overlain by gray and mottled limestone, and this in turn overlain by dark-gray limestone with irregular calcite veinlets	50
9	Shale, tan, thin-bedded. A sandy limestone in middle contains hematite pseudomorphs after pyrite	35
8	Limestone, vermicular (Bluebird type)	40
7	Dolomite, white, laminated (Lynch dolomite type). Vermicular limestone of Bluebird type in middle with some black chert nodules	100
6	Limestone, vermicular (Bluebird type). More massive bedded at top than bottom	175
5	Limestone edgewise conglomerate with some beds of mottled or banded limestone and also some beds of gray shale with ½-inch concretions	30
4	Shale, olive-gray, splintery, interbedded with thin blue-gray mottled limestone	100
3	Limestone, dark-gray to black with conspicuous irregular mudstone bands and mottles of lighter tan. Blotchy in part. Some beds have sandylike weathered surface (Hartmann type)	150
2	Shale and platey limestone, generally light gray-blue. Some limestones have yellowish blotches. A porous black sandstone probably near bottom	425
1	Limestone, dark gray-blue, banded, mottled, and lumpy (Hartmann type). Large oölites and pisolites in lower part	100
	Conformable on Ophir shale	
	Total thickness	1375±

Certain characteristic and widespread lithologic features of the Cambrian beds in the Utah region have been described by Gilluly (1932, p. 12–20). Three are very common, and he named one of them the "Bluebird type." The other two will be referred to as the "Hartmann type" and the "Lynch type" in this report. The Bluebird type is a dark-gray crystalline dolomite or limestone "with a spangling of white rods and vermicular tube-like markings which average perhaps one-sixteenth of an inch in diameter and are a quarter to half an inch in length" (Gilluly, 1932, p. 16). The type locality of the Bluebird dolomite is in the Tintic district.

The Hartmann type is characteristic of the Hartmann formation in the Ophir district. It is a "limestone banded with mudstone material which is distributed in a peculiar manner so that the weathered surfaces of the limestones are irregularly mottled with light gray, buff, and brown. Some of the mudstone bands are parallel to the bedding and are connected at closely spaced intervals by streaks of the same material normal to the bedding and suggestive of filled mud cracks." Other strata have mudstone bands "arranged in extremely sinuous, bulbous, biscuit-shaped, or pseudoconcretionary patterns" (Gilluly, 1932, p. 12, 13). Beds of oölitic and pisolitic limestone are commonly interbedded.

The Lynch type is taken from the Lynch dolomite of the Ophir district. It is a finely laminated gray to white dolomite which weathers white and is a very conspicuous marker in the field. The Lynch dolomite formation is one of interbedded dolomites of the Lynch type and the Bluebird type. The lithologic types occur at several different horizons in the Middle and Upper Cambrian sequences and, therefore, cannot be used for close correlation.

The writer has constructed columns for the northern Utah area (Richardson, 1913 p. 407; Walcott, 1908, p. 191; Deiss, 1938, p. 1105–1124), for the area of this report, and for the Ophir district (Gilluly, 1932, p. 12–20) and finds an excellent agreement in lithology and sequence with the Ophir section but several discrepancies between it and the northern section. The section exposed in Durst Mountain probably represents the Hartmann formation, the Bowman limestone, and part of the Lynch dolomite of the Ophir district, but its correlation with the northern Utah section (Langston, Spence, Ute, Bloomington, Nounan, St. Charles) is doubtful.

Gilluly has assigned the Hartmann and Bowman formations to the Middle Cambrian and the Lynch dolomite to the Middle and Upper Cambrian.

At the top of the Cambrian limestone sequence in Ogden Canyon (above the Hermitage) is a fairly fine-grained, medium-gray, uniform dolomite or dolomitic limestone which may be the Laketown dolomite. Doctor Williams and the writer searched for fossils but found none. This unit is included in the Upper Cambrian limestones.

DEVONIAN (?) SYSTEM

Three Forks (?) formation.—Resting upon the white laminated dolomite of the Cambrian is a group of beds composed mostly of sandstone and shale. They crop out only in two places in the area of this report—*viz.*, Durst Mountain and Ogden Canyon. On Durst Mountain the best exposure is along the crest of the ridge between Devil's Hollow and Big Hollow where 600 feet of beds was measured. The sandstones are thin-bedded and inconspicuously tan and red. The lower quarter of the formation is made up mostly of drab shale, and above this some gray quartzitic sandstone beds occur. A conspicuous group of beds in the upper part is conglomeratic. Small red pieces of crumpled shale, $\frac{1}{4}$ to 3 inches long, are embedded in a calcareous sandy matrix. Float indicated the presence of some thin limestone beds. No fossils were found, but, as no extended search was made, they may be present.

The reddish soft-weathering outcrop could be traced northward across Devil's Hollow and Spring Hollow whence it passes just east of Durst Peak into Miller's Gulch and Durst Canyon. The exposure terminates southward against a fault. The beds at the head of Miller's Gulch along the crest of the high ridge extending eastward from Durst Peak are principally sandstone and sandy and shaly limestone and dolomite. The sandstone is tan, yellow, or red, and the limestone is pink and yellow, mottled, lumpy, and thin-bedded. A whitish-weathering limestone 100 feet thick with calcite veinlets is about in the middle of the formation, and a gray fossiliferous limestone with irregular chert nodules is at the bottom. The section here was estimated to be 775 feet thick.

In the Paleozoic section in Ogden Canyon a tan slope and talus-making group of

beds about 500 feet thick occurs just below the fossiliferous Madison limestone which in turn underlies a phosphatic shale and limestone unit, most probably the base of the Brazer. No fossils were found in the tan slope-making unit in Ogden Canyon. It probably is the same formation as the nonfossiliferous group of tan and red beds of Durst Mountain, however, because it occupies the same stratigraphic position.

Time was not taken to study the Paleozoic limestones at the head of Hardscrabble and City Creek canyons. They are slightly south of the area of this report but may contain the Upper (?) Devonian beds to which Gilluly (1932, p. 22) refers in citing an oral communication of Edwin Kirk and A. A. L. Mathews.

Blackwelder (1910a, p. 528–529) mentions 250 feet of red and pinkish shales and thin-bedded limestones on the west slope of Durst Mountain and thinks they may be Devonian. These beds are probably part of the formation that the writer has described above.

The chief reason for assigning the 600 feet of tan and red sandstones and drab shales to the Devonian is Richardson's work in the Randolph quadrangle, Utah, just a few miles to the north. He correlates rocks with similar lithologic characteristics at this horizon with the Devonian Threeforks and Jefferson formations (1913, p. 411–412). The Threeforks is described as a soft formation of impure reddish limestone about 200 feet thick, and the Jefferson as a dark dolomitic limestone. In the north-central Wasatch area the lithology of the beds in question is more characteristic of the Threeforks than the Jefferson, and hence the formational name, Three Forks, will be used provisionally.

MISSISSIPPIAN SYSTEM

Madison limestone.—Approximately 1750 feet of Mississippian strata was measured in the Durst Mountain area, of which 650 feet is provisionally assigned to the Madison formation and 1100 feet to the Brazer. The presence of two units of distinctive lithology induced the writer to subdivide the sequence of beds heretofore spoken of as Mississippian beds (Blackwelder, 1910a, p. 528–530), and this subdivision has recently been confirmed by Williams (1943, p. 596–599).

The following details of the Madison were obtained along the crest of the high ridge at the head of Miller's Gulch east of Durst Peak.

Units		*Thickness (Feet)*
3	Limestone, dark gray, weathers to sandylike surface. A gray limestone with light-gray chert nodules about in middle of unit. Calcite veinlets through limestone. Fossiliferous. About	400
2	Limestone, pinkish, argillaceous	50
1	Limestone, dark gray, dolomitic. Contains a 2-foot iron oxide vein and a barite vein on high ridge at head of Miller's Gulch	200
	Approximate thickness	650

Williams gives the thickness of the Madison in Weber Canyon, at the south foot of Durst Mountain, as 755 feet.

The Madison was estimated to be 500 to 800 feet thick in Ogden Canyon.

Brazer formation.—A group of limestone, shale, and sandstone beds just below the red sandstone of the Morgan formation is assigned to the Brazer. Its units are

fairly distinct from the underlying Madison and together serve as a convenient formation to map. The succession is best exposed on the north side of Round Valley and on down the canyon to the quarries at the roadside just northeast of Morgan. The major units from top to bottom are:

(*Units*)		*Thickness (Feet)*
5	Limestone, tan-cream, gray, and carbonaceous black, mostly lithographic. White chert nodules abundant in beds from top to bottom of unit. Large cup corals, productids, and spirifers plentiful. Approximately	150
4	Limestone, black finely crystalline. Beds up to 18 inches thick. Much microscopic fossiliferous material, brachiopods, and biostroms of bryozoans	50
3	Shale and sandstone. Shale purple and olive tan; sandstone tan. Some light-gray limestones probably interbedded	100
2	Limestone, medium gray to creamy, finely crystalline. Many large horn and colonial corals	200
1	Sandstone, tan to red. Calcareous with some light-gray quartzitic beds. A few limestone beds intercalated. Approximately	600
	Total approximate thickness	1100

Unit 1 was mapped separately to show the small anticline in the Madison and Brazer limestones at the town of Morgan. Units 5, 4, 3, and 2 probably correspond to units 5 and 4 of Williams' recent paper (1943, p. 598). He lists a typical Brazer fauna from these units. His thickness of 1270 feet for his lower two units (3 and 2) is double that of this report's corresponding unit 1, and the writer has not been able to check the measurement in the field.

Units 3, 4, and parts of 2 seem to be similar to Upper Mississippian units described by Blackwelder (1910b, p. 549) in Dry Bread Hollow, a tributary of the South Fork of Ogden River about 20 miles north of Morgan.

Interbedded limestone and thick hard gray sandstone beds in Ogden Canyon are believed to be part of the Brazer formation. They crop out just above the Pineview Dam along the highways on each side of the dam and may be seen along the abandoned road below the dam. The limestone is gray and black, and some beds carry black chert nodules. Below this series in Ogden Canyon is a 20-foot unit of black phosphatic shale and limestone (Blackwelder, 1910b, p. 539) which, according to Doctor Williams, is probably the base of the Brazer there.

PENNSYLVANIAN AND PERMIAN SYSTEMS

Morgan sandstone.—The type locality of the Morgan sandstone is in Round Valley east of the town of Morgan where it lies below the type Weber quartzite and above the Brazer limestones. The name "Morgan sandstone" was first used by Blackwelder (1910a, p. 529), but he states that F. B. Weeks in an unpublished manuscript called this unit the "Morgan formation." Blackwelder's description follows:

"The prevailing rock . . . . is earthy sandstone, which is relatively soft. Fresh surfaces are generally white or pink, but on exposure the rock turns brick red. Some beds are distinctly shaley, and here and there thin layers of gray limestone with a few fossils are interbedded with the series. The Morgan formation passes upward through alternating gray shales, limestones, and sandstones into the more or less calcareous base of the Weber quartzite. In this transition zone there are a few fossils—chiefly Lingulas and Discinoids. The lower limit of the formation is sharp, for the earthy red sandstones rest upon a cavernous weathered surface of fossiliferous gray limestone. Just above the contact lies a coarse sandstone which consists of well-rounded frosted sand grains bound

in a deep red matrix and including bits of limestone and black chert from the underlying series . . . . the relations here clearly indicate an unconformity. . . ."

The fauna found in it by Blackwelder suggests an early Pennsylvanian age which is confirmed by Williams' recent work (1943, p. 614). The Morgan sandstone is approximately 1000 feet thick in Round Valley.

Weber quartzite.—The type section of the Weber quartzite is in upper Weber Canyon east of Morgan where it was named by King (1876, p. 477–479). No section of the formation is given by him, but the thickness was estimated to be 5000 to 10,000 feet. The name "Weber" has since been extended as far east as central Colorado, and many references have been made to the type locality, but no description has ever been published. The writer encouraged Kenneth McGlone, who was assisting him in 1936, to measure and describe the Weber section. His section follows:

Units		*Thickness (Feet)*
13	Quartzite, white, cliff maker. Joint surfaces have a rusty glazed appearance. Upper part of unit is brecciated and weathers cavernous	205
12	Quartzite and sandstone. Unit mostly covered	320
11	Quartzite, grayish white. Pitted surface	10
10	Covered, probably quartzite	145
9	Chert, grayish white, mottled, appears brecciated. Under the microscope the rock is 90 to 95 per cent chert. Sand grains of quartz, rutile, tourmaline, and zircon were seen. Rhombic casts of dolomite and veinlets are filled with secondary silica	3
8	Quartzite, light gray, medium gray, and buff. Forms high cliff in upper part and steps in lower. Capped with a dark-blue cherty limestone. Under the microscope the rock is a sandstone with siliceous cement. The grains range from .05 to .9 mm. in diameter. The larger grains are rounded, and the smaller ones are subangular to angular. Minerals in minor amounts are chert, ilmenite and leucoxene, microcline, tourmaline, and zircon	185
7	Mostly covered. Float is chiefly sandstone, reddish brown, buff, and light gray. At top is a cliff of buff to white quartzite	325
6	Quartzite, light gray and white. A thin dark-blue sandy limestone occurs 75 feet above the base. A limey sandstone at the top of the unit is fossiliferous. Microscopically the calcareous sandstone is about 15 per cent calcite, and the grain size varies greatly and averages about .15 mm.	265
5	Quartzite, gray to brown, cliff maker	100
4	Quartzite, grayish white to buff and reddish brown. Medium- to fine-grained. Unit is dominantly a slope maker	480
3	Limestone, sandstone, and quartzite interbedded. Limestones dark gray with abundant chert nodules but few fossils. Three limestone beds found, one at base, one 240 feet up, and another at the top of unit. Sandstone has calcareous cement (20–30 per cent), and quartzite is a siliceous sandstone. Grains .09 to .11 mm. in diameter and subangular. Fossils were found near the base in float	355
2	Sandstone, light gray at base and buff to brownish red above. Calcareous. The brownish-red sandstone has an iron-oxide cement and a porosity of 20 to 25 per cent. The grain size is uniform and averages about .08 mm. The accessory minerals are ilmenite and leucoxene, microcline, rutile, tourmaline, zircon, sericite, and chlorite. Cliffs of quartzite at 200 and 250 feet from base	605
1	Sandstone, light gray to white, calcareous, medium- to fine-grained. One bed of nonfossiliferous dark-blue limestone with chert concretions bear middle. Calcite sand grains present and much larger than the quartz grains (.07 to .10 mm.). In thin section studied about 40 per cent of rock was calcareous cement	125
	Total thickness	3000±

This thickness justifies the doubt that Blackwelder (1910a, p. 531) registered regarding King's figure of 5000 to 10,000 feet. Fossils collected by Blackwelder (1910b, p. 540) were considered by Girty to be of Pennsylvanian age, and Williams' (1943, p. 615) collection proved to be without more precise stratigraphic value. However, Williams considers the age of the Morgan sandstone fairly well established, believes it grades upward into the Weber, recognizes the Weber as representing a short part of geologic time, and therefore concludes that the Weber is also lower Pennsylvanian in age.

None of the quartzites are metamorphic rocks, according to McGlone.

Park City formation.—Approximately 1600 feet of beds above the typical Weber quartzite and below the dark brick-red Woodside shale contains considerable tricalcium phosphate. Their outcrops in Weber Valley above Morgan have been visited and discussed by geologists studying the western phosphate deposits and the problems of the Permian and Pennsylvanian stratigraphy during the past 40 years, and a number of publications concerns them directly or indirectly. The most recent papers are those of Williams (1939, p. 82–100), Baker and Williams (1940, p. 627–629), Williams and Hanson (1942, p. 8–14), and Williams (1943, p. 597, 617).

For convenience the section in Weber Canyon has been divided into three divisions. The upper division is about 720 feet thick and is made of light-gray to dark-gray limestones, gray to olive to purplish shales, and some calcareous sandstones. At several horizons dark chert nodules occur. Fossils were noted at several horizons. The lower limestone beds of this division may be slightly phosphatic. The middle division is 615 feet thick, according to Williams and Hanson (1942, p. 12–13), and carries the commercial phosphate beds. Their detailed section of the middle division, with minor additions by the writer, is as follows:

Units		*Thickness (Feet)*
8	Covered, probably shale with a small phosphate content	25
7	Siltstone and calcareous siltstone, dark brown, $Ca_3(PO_4)_2$—15.7%	3
6	Oölite, brown shale, and siltstone. Oölitic beds range from 27–32% $Ca_3(PO_4)_2$	12½
5	Sandstone, light gray calcareous; siltstone fuscous, black	35
4	Oölite	1
3	Limestone, dark gray, dense, with great masses of dark chert. Approximately	230
2	Limestone, sandy limestone, and siltstone, dark gray	300
1	Shale, fuscous, black	8
	Total	615

The lowest division is a 270-foot sandstone that weathers dark gray. It is made up of calcite and quartz grains in about equal proportions (average diameter 0.05 to 0.07 mm.) cemented by calcite.

The upper two divisions appear to be the equivalent of the Phosphoria formation in northern Utah (Williams, 1939, p. 12), and all three divisions together probably represent Boutwell's Park City formation in the central Wasatch (Williams, 1943, p. 618). The upper two divisions of the Phosphoria section in Weber Canyon are Permian in age, but the age of the lower one is not yet definitely fixed.

TRIASSIC SYSTEM

Woodside shale.—The Woodside shale is dark brick red and sandy. Some beds are micaceous, and a few beds of red sandstone an inch thick are interbedded in the shale. Ripple marks are common throughout. It represents a rather sudden lithologic change from the underlying Park City limestone beds, but local outcrops yielded little evidence of the nature of the contact. The best exposure found is in a railroad cut in upper Weber Canyon above Morgan.

The type section of the Woodside is in the Park City district (Boutwell, 1907, p. 446–452). It ranges in thickness in the central Wasatch from 800 to 1000 feet (Mathews, 1931, p. 5) but evidently thins northward to Weber Canyon where it is about 600 feet thick.

Thaynes formation.—The Thaynes formation in upper Weber Canyon is made up chiefly of dark-gray, medium-gray, and tan-gray limestone that is fine-grained to fairly crystalline. Some of the beds are 3 feet thick. Several drab-gray shale members are intercalated, and near the top is a group of olive and red splintery shale beds. Ripple marks are common throughout. The section here has a thickness of approximately 2200 feet.

Mathews, who has studied the Thaynes formation in central Utah (1931, p. 6–36), finds an abundant fauna in several zones and divides the section into two formations. The lower formation, the Pinecrest, is Lower Triassic, and the upper, the Emigration, is Middle Triassic. These two divisions probably occur in Weber Canyon, but no special search was made to locate their contact. Mathews' total section measured 1615 feet in comparison with 2200 feet for the Weber Canyon section.

Ankareh formation.—The Ankareh in upper Weber Canyon is estimated to be 1100 feet thick. It is made up mostly of shale although sandstone beds up to 4 inches thick are common. The shale and sandstone beds are dull red, dark purple red, and gray green. Some beds that appear to be 2 to 4 feet thick are shale that breaks down into splintery fragments. The formation is ripple-marked throughout. It may be distinguished in small outcrops from the red Woodside shales by the presence of gray-green shales intercalated in the red.

The type Ankareh is in the Park City district (Boutwell and Woolsey, 1912, p. 58), where it is 1150 feet thick. Mathews (1931, p. 37–40) believes that the Ankareh is Upper Triassic. Previously the formation has been assigned both a Jurassic age (Reeside, 1929, p. 47–63) and a Triassic age (Branson, 1927, p. 607–630).

JURASSIC SYSTEM

Nugget sandstone.—The Nugget sandstone is well exposed in upper Weber Canyon just below Devil's Slide. It is a massive, cross-bedded, cliff-making, pink to salmon sandstone. Mathews (1939, p. 42) reports a basal conglomerate in this section, but it was not seen by the writer. The formation is about 1500 feet thick.

The type Nugget is in southwestern Wyoming at Nugget Station (Veatch, 1907, p. 56) where the formation is 1900 feet thick. Only the basal 600 feet is exposed in the Park City district (Boutwell, 1912, p. 59). It is probably Lower or Middle Jurassic (Reeside, 1929, p. 47–63; Mathews, 1931, p. 42–43).

Twin Creek formation.—The gray shale and limestone beds of the Twin Creek

formation were observed by the geologists of the Fortieth Parallel Survey but were simply called the Jurassic shales and limestones. Veatch (1907, p. 56) described the Twin Creek formation of southwestern Wyoming, and since then the characteristic exposure in Weber Canyon (Devil's Slide) has been spoken of as the Twin Creek.

The outstanding units exposed in the Devil's Slide area from top to bottom are as follows:

Units		*Thickness (Feet)*
10	Shale, tannish gray, calcareous, platey. Considerable limestone near base. Two red shales 375 feet up from base of unit. Certain shales, especially near top, break up into large splintery fragments. Top covered by Wasatch	700
9	Limestone, cream to dark gray, massive. Lithographic in part with interbedded gray shales. These limestones are the "high rock" of the cement rock quarry, in contrast to the "low rock" of the shales of the base of unit 10	400
8	Shale, greenish-gray	50
7	Limestone, dark gray, and gray with reddish and yellow blotches. Crystalline in part and lithographic in part	200
6	Shale, gray, calcareous, splintery (cement rock)	300
5	Shale, gray, and limestone, argillaceous, platey	700
4	Limestone, tannish cream, lithographic. Darker gray toward top with argillaceous platey limestone	150
3	Shale, gray	60
2	Devil's Slide limestone, gray lithographic	30
1	Mostly shale, unit poorly exposed. Perhaps several gray limestones within unit. In the central Wasatch two rather thick oölitic limestones occur in this unit (Mathews, 1931, p. 44), with a varied Jurassic fauna 20 feet above the basal 60-foot oölitic limestone	550
	Total thickness exposed in Weber Canyon	3140±

In Toone Canyon, 8 miles up Lost Creek from Devil's Slide, there is approximately 600 feet of very splintery, light-gray shale with a brownish-gray arenaceous limestone at the top. The brownish limestone is overlain by dull-red beds provisionally called the Entrada, and hence the arenaceous limestone is assumed to represent the top of the Twin Creek formation. The top of the Twin Creek is not exposed at Devil's Slide, but the splintery character of the shale in Toone Canyon seems distinctive of the upper division of the Twin Creek as exposed near Peoa on the Weber 25 miles to the south, and hence the 600 feet of shale in Toone Canyon is thought to be above the Devil's Slide section.

The type Twin Creek is 3500 to 3800 feet thick. The formation exceeds 3140 feet in upper Weber Canyon, and if 600 feet of beds should be added to unit 10, as suggested by the exposure in Toone Canyon, the total thickness would be 3700±. The similarity in thickness of this section to that of the type section in southwestern Wyoming was unexpected because the formation is only about 1200 feet thick at Peoa 25 miles southward, where both the base and top are exposed and no faults within the formation have been recognized.[2] Mathews (1931, p. 44) estimates about 2000 feet of Twin Creek in Parleys Canyon of the central Wasatch.

Of interest in the lower part of the Peoa section is a light-green sandy bed com-

[2] Plane-table measurement by Hyrum Schneider, University of Utah.

posed of 90 per cent chlorite and 5 per cent each of biotite and calcite. No glass was noted, but it seems probable that it is a thoroughly altered tuff. Mansfield (1927, p. 97) describes a light-green tuff in the base of the Twin Creek in southeastern Idaho. This unusual lithology suggests that the same bed extends from one locality to the other.

Entrada (?) and Curtis (?) formations.—South of the area of this report, between Peoa and Wanship, 4000 feet of variegated beds crops out above the Twin Creek. Nowhere within the area mapped is a complete section exposed, and some of the beds are not exposed at the surface anywhere within the area. The beds have been measured and tentatively subdivided by Hyrum Schneider, and the provisional formational names, Entrada and Curtis, have been applied by him on the basis of observations farther south in company of A. A. Baker who is mapping the Strawberry quadrangle. The upper part of the sequence has been traced to the type locality of the Kelvin conglomerate in the central Wasatch, described under the next heading. Possibly these names will not hold upon the completion and publication of work by Baker, in the area mentioned, and E. M. Spieker in central Utah, and therefore their provisional character should be emphasized. Thick and coarse conglomerates in the upper part of the sequence mark the first important diastrophism of the general Sierran-Laramide revolution in the area, and it is therefore urgent that the ages of the formations be determined; but it is best at present to venture little in the way of correlation. After reliable stratigraphic correlations have been made, however, the early diastrophic stages in the north-central Wasatch may better be related to those of other parts of the Rocky Mountains.

The following descriptions are those of the writer and correspond chiefly to field notes of Schneider.

Entrada (?) formation. Deep, dull-red shale and siltstone units that weather into a soft red soil are above and below a more resistant middle dull-red sandstone and shale unit. The sandstone beds range up to 8 inches thick, and some are finely speckled with white and gray grains. Thickness, 1170 feet.

Curtis (?) formation. A lower thin-bedded shale unit is gray to dark gray and mostly covered; a middle sandstone unit is tan and greenish tan and is speckled with light and dark grains. Some limestone float was noticed. An upper unit of limestone is about 6 feet thick, gray, slightly lumpy, and rough with algal (?) banding. This upper limestone may belong to the overlying Kelvin (?) formation. Total thickness, 150 feet.

The Peoa-Wanship section on the Weber River was visited by Stanton (1893, p. 44–45) in 1891 or 1892, and, although the nomenclature used in the present report existed only in part at that time, his published units correspond to the Nugget (Stanton's unit 10), the Twin Creek (units 9 and 8), the Entrada (unit 7), the Curtis (unit 6), and the Kelvin (units 5 and 4). He correlated the Entrada, Curtis, and at least the lower division of the Kelvin (?) with Veatch's Beckwith formation (Veatch, 1907, p. 57–59).

Stanton found some of these formations in Chalk Creek, just east of the area of the present report and also in the upper part of East Canyon, just south of the area mapped. Besides these near-by exposures the writer has recognized the Twin Creek,

Entrada (?), and the conglomerates of the Kelvin (?) in Toone Canyon, a tributary of Lost Creek, and has identified the upper beds of the Kelvin (?) in the core of the Coalville anticline.

A syncline at the mouth of Toone Canyon contains the uppermost sandy limestone bed of the Twin Creek, the red Entrada (?) shale, and the lowermost shale of the Kelvin (?). An anticline up the canyon from the syncline exposes the upper Twin Creek beds, and on the east flank the contact of the Twin Creek and Entrada (?) is clear, but no beds to represent the Curtis were seen. A 50-foot interval at the base of the Kelvin (?) is covered, and therefore a few Curtis beds may be present. The conglomerate-bearing beds of the Kelvin have approximately a vertical dip and are concealed by the overlying Wasatch beds at a point in Toone Canyon about 2½ miles above its mouth. Beds of the Wasatch group (to be described shortly) have been arched up in a broad anticline and overlie the truncated edges of the folded Jurassic beds.

Stanton found a *Trigonia quadrangularis*? in beds near Peoa that seem to be the Curtis (?), as well as can be made out from his writing. The formational names of the San Rafael Swell, Entrada, and Curtis are used in this report, but it possibly would be just as proper to designate the lithologic units by the names Preuss and Stump formations of southeastern Idaho (Mansfield, 1927, p. 98–99).

CRETACEOUS SYSTEM

Kelvin (?) formation.—The name is used very provisionally to include two sequences of strata, between Peoa and Wanship, a lower shale with four conspicuous groups of conglomerate beds and an upper variegated shale and sandstone. Between the conglomerates of the lower sequence is chiefly shale with interbedded sandstone. The shale is salmon red, pinkish white, purplish red, and brick red, and above the lowest conglomerate it is probably bentonitic. The sandstone is light gray and pinkish gray, usually massive and cross-bedded. The pebbles and boulders in the lower conglomerate are all chert in black, red, and yellow colors, some banded and some not. Many are polished like gastroliths. A few quartzite pebbles that may be Weber quartzite occur in the second conglomerate. In the upper coarse conglomerate exposed in the road cut most of the large boulders are quartzite. The small pebbles are mostly chert. Some Nugget (?) sandstone and a few Paleozoic limestone pebbles were seen. The conglomerate-bearing division is about 700 feet thick.

In the upper thick interbedded sequence of shale and sandstone, the shale ranges from light gray to purplish gray to pink to brick red, and the sandstone from light gray to light brick red. The sandstone is commonly coarse-grained, massive, and cross-bedded. The lithology and thickness of the unit probably changes considerably toward the west in the drainage area of Parleys and Emigration creeks. The thickness is about 2300 feet in the exposure between Peoa and Wanship.

In Emigration Canyon about 25 miles west of the exposure at Peoa on the Weber River, a thick conglomerate, sandstone, and shale sequence was named the Kelvin conglomerate by Mathews (1931, p. 48). R. E. Marsell and the writer have traced the conglomerate-bearing beds in lower Emigration Canyon eastward across upper East Canyon into the Wanship-Peoa area and recognize the lower conglomerate unit

of the section here described as equivalent, at least in part, to Mathew's Kelvin conglomerate. The upper unit of variegated shale and sandstone is believed by Stokes (1941, p. 63) also to be equivalent to the Kelvin conglomerate. He points out that the conglomerate beds become coarser and more numerous toward the west front of the Wasatch and believes that this indicates a western highland as a source. He further observes that most of the chert boulders and pebbles are from the Permian and Pennsylvanian formations which must have underlain large parts of the highland at the time.

As the upper unit on the Weber below Peoa is not conglomeratic the term Kelvin formation is used rather than Kelvin conglomerate. No fossils have been found in the Kelvin formation, and the few scattered outcrops both north and south of the Peoa section have not been correlated. Therefore, its age is not known. Mathews erred in calling certain beds above the mouth of Lambs Canyon in Parleys Canyon Morrison, according to Spieker and Schneider (personal communication); the beds there are the Entrada (?) and Curtis (?) of the Peoa section. Possibly the Kelvin is Morrison in age, or possibly equivalent to the lower part of Spieker's Indianola group in central Utah, which he believes is lower Colorado, or to Mansfield's Ephraim conglomerate (1927, p. 103–104) of the Gannett group, in the southeastern Idaho, which is possibly Lower Cretaceous.

Aspen and Frontier formations.—A thick Cretaceous sequence above the Kelvin formation on the Weber River at Rockport has been visited and discussed by many geologists. Stanton's (1893, p. 44) study and Veatch's (1907, p. 103) correlations as a result of work in southwestern Wyoming contain a thorough review of previous work and incorporate considerable new paleontologic information and a generalized stratigraphic column. The names Bear River (?), Aspen, and Frontier formations were applied to the Rockport section in these reports because of lithologic and paleontologic similarity to the near-by type sections of southwestern Wyoming. The section at Coalville on the Weber received even more attention than the Rockport section. The Coalville coal field was reported on by Wegemann in 1915, who gave a more detailed section than Stanton and discussed the correlation problems of those upper units that are nearly conformable with the Almy conglomerate in Echo Canyon.

In the core of the Coalville anticline (Pl. 1) is exposed about 850 feet of variegated shale and white and gray sandstone which undoubtedly are the upper part of the 2300-foot section of Kelvin (?) beds in Weber River Valley north of Peoa. These lowest beds in the Coalville anticline are Wegemann's unit 16 (1915, p. 163), of uncertain age. Monnett has begun a study of the formations exposed at Coalville and Peoa and finds that about 600 feet of black shale containing fish scales overlies the variegated Kelvin (?). The black shale with fish scales is generally regarded as characteristic of the Aspen formation. Associated with the black shales in the Peoa section are some bentonite and porcellanitelike beds, also common in the Aspen farther north. As these beds rest directly upon the variegated shale and sandstone, it is concluded tentatively that no Bear River beds are present in either the Coalville or Peoa sections, and also that the Aspen is thinner than it is farther northeast.

Above the Aspen formation is a coal-bearing series of hogback-making gray sand-

stone beds separated by dark shale and friable sandstone, altogether about 10,000 feet thick, according to Monnett's measurements. This formation includes Wegemann's units 3 to 13 plus a large part of his unit 14 and the lower 1500 feet of his unit 2. The sandstone hogback makers contain characteristic Frontier fossils (Stanton, 1893, p. 44; Veatch, 1907, p. 103). Wegemann's map of the area shows no divisions of this thick formation except the positions of three main coal beds, but in order to delineate the structure better on the map of the present report the ridge-forming sandstone beds were mapped separately. They are numbered 1 to 5. The Dry Hollow coal bed of Wegemann is near the base of No. 4 sandstone, the Wasatch coal bed is between sandstones 2 and 3, and the Spring Hollow coal bed is near the top of sandstone 1.

Wegemann describes a coarse conglomerate as resting in places directly on sandstone No. 3 and in others as separated from it by 180 feet of shale. The conglomerate is brown, contains pebbles principally of limestone and sandstone, the largest of which are 6 inches in diameter, and is 40 to 100 feet thick. He also mentions numerous thin beds of conglomerate and coarse sandstone at other stratigraphic positions. These conglomerates suggest that the western highland was being elevated during most of Frontier time.

The bentonite and porcellanitelike beds in the Aspen and another thick bentonite bed about 5000 feet above the base of the Frontier (personal communication, V. B. Monnett) mark times of volcanism, but the sites of eruption must have been rather distant.

MONTANA OR PALEOCENE SERIES

Henefer formation (new name).—Stanton (1893, p. 39) mentioned "room for 1500 feet of strata" above his No. 10 unit [top (?) of the Frontier] and below the "Echo Canyon conglomerate." Wegemann's (1915, p. 163–164) unit 2 includes the same strata, but as he lists a thickness of 2500 feet some of the upper shale and sandstone beds of Stanton's unit 10 may be included also. The writer computed a thickness of 1250 feet of beds below the Pulpit conglomerate of Echo Canyon and above the highest gray sandstone hogback maker that divides Carruth Canyon from Lewis Canyon (Pl. 1) and crops out just north of the railroad in Grass Valley Canyon. The hogback maker is truncated by the overlying tan, gray, and pink slope-making succession of sandstone and shale beds at the head of Lewis Canyon and is therefore probably part of the Frontier. The pink and gray series disappears under the coarse Almy conglomerate about a third of a mile above Echo Dam but reappears down Weber Canyon above the mouth of Franklin Canyon on the other side of a syncline. From Franklin Canyon northwestward the beds form both sides of Henefer Valley to the narrows of the upper Weber Canyon. The wide valley of Little East Canyon is eroded in an anticline of these beds, and they are also well exposed in Harris Creek Canyon northeast of Henefer. It seems appropriate, therefore, to name the beds the Henefer formation.

The Henefer formation is 1250 feet thick at Echo dam, and in Harris Creek Canyon only 7 miles away 10,000 feet of beds is exposed. The Frontier formation does not crop out anywhere in the core of the Henefer anticline, and therefore the Henefer formation is thicker than the 10,000 feet of beds exposed.

In Harris Creek Canyon the formation consists of red, pink, purple, and light-gray shale, red and gray sandstone of various textures with numerous lenses of pea and pebble conglomerate, and gray, tan, brown, and red pebble and boulder conglomerate. A coarse conglomerate that forms a ledge at the highway just east of the bridge at Henefer is made up of 75 per cent of pebbles and boulders from the Twin Creek, Nugget, Ankareh, and Thaynes formations. The other 25 per cent is quartzite pebbles mostly from the Weber. A few limestone pebbles may be from the Paleozoic formations. In Little East Canyon in the axial zone of the anticline is a 200-foot bed of soft light-gray sandy tuff and lumpy limestone. According to Hatch (personal communication) the tuff is of rhyodacitic composition (Johannsen's classification). The tuff and limestone beds are overlain by soft red shale and underlain by interbedded red shale and sandstone, and gray pea conglomerate.

If Hayden's (Hayden, 1869, p. 90; Veatch, 1907, p. 87) designation of the type locality of the Wasatch group be taken literally—*viz.*, route from Carter, Wyoming, down Echo Canyon to the Weber, and down the Weber to the narrows below Henefer—then the Henefer formation is the lowest division. The beds were not considered part of the Wasatch, however, by Wegemann (1915, p. 164), who suggested that they correlate with the Hilliard formation of Veatch's section in southeastern Wyoming. Wegemann reports fossil leaves and fresh-water shells but says the fossils were too poor to identify. Fossil plants were collected by C. A. Arnold and the writer, in a small quarry just east of Henefer on the main highway. The beds cropping out in the quarry stand on end and are about in the middle of the thick section. According to Arnold the plants probably represent a single new species and cannot be used for age determination.

Three formations occur in southwestern Wyoming between the Frontier formation and the Wasatch group—the Hilliard, Adaville, and Evanston. The Henefer formation of Weber Valley is between the Frontier and Wasatch and therefore probably is equivalent to part or all of the Hilliard, Adaville, and Evanston. The major unconformity in the Wyoming section is at the base of the Evanston, and the Henefer is unconformable on the Frontier of the north-central Wasatch. Both are overlain unconformably by the conglomerate beds of the Wasatch group. Therefore, if the Henefer is equivalent only to one of the three Wyoming formations it seems most likely that it would be the Evanston.

The Evanston formation differs from the Henefer in having carbonaceous shale and no red beds (Veatch, 1907, p. 77) but is similar in having brown conglomerate beds and numerous plant fossils at different horizons. If these formations were thoroughly worked over for fossil leaves reliable correlation might be made. The Evanston is considered lowermost Paleocene (Puercan) (Wood *et al.*, 1941, Pl. 1, p. 1).

Walton (1944, p. 117–120) has recently described a new formation, the Currant Creek, in the western part of the Uinta basin, about 50 miles southeast of the Henefer Valley, and it appears to be about the same age as the Henefer formation. The Currant Creek formation consists of conglomerates, sandstones, and variegated shales which, in the western part of the Uinta Basin, are 4500 feet thick, transect the Mesaverde (Niobraran) beds, and unconformably underlie the Eocene strata of probable Uinta age. No fossils have been found in the formation, but on the basis of its lithology and stratigraphic relations Walton believes it may be equivalent in

all or part to the Evanston and Almy (p. 120) of southwestern Wyoming, and in his correlation diagram (Walton, 1944, Pl. 3) he suggests the Evanston as the major equivalent. He points out also that the Currant Creek formation probably correlates with the Price River and North Horn formations of the Wasatch Plateau. These two formations span uppermost Cretaceous and lower Paleocene time.

The materials of the Currant Creek formation were derived from mountains immediately to the west. Large alluvial fans were deposited along the piedmont, while farther east on river flood plains sands and muds were spread out. The Henefer formation probably also came from the west, but no study has yet been made of possible areas of outcrop east and northeast of Henefer Valley beyond the limits of the present report, so little can be said at present of its distribution and source.

PALEOCENE SERIES

Almy conglomerate.—The coarse, red, cliff-making conglomerate in the lower 15 miles of Echo Canyon was first described by Hayden (1869, p. 90). He correlated it with beds later called the Almy formation of southeastern Wyoming by Veatch (1907, p. 89) who believes this relation correct. The Almy formation in Wyoming is the lowest of three formations—*viz.*, the Almy, Fowkes, and Knight—that make up the Wasatch group (Veatch, 1907, p. 88).

The Almy conglomerate in lower Echo Canyon is divided into two divisions by a conspicuous 5- to 30-degree angular unconformity. The erosion surface is seen best in Saw Mill Canyon, 4½ miles up Echo Canyon from the Weber. The overlying conglomerate and sandstone beds can be traced southeastward across the truncated edges of the underlying conglomerate and over the divide into Grass Valley where they rest on the Henefer formation. On the other hand, the two divisions of the conglomerate become concordant to the northwest and northeast and were not differentiated.

The upper division of the Almy conglomerate is here named the Saw Mill conglomerate from Saw Mill Canyon, and the lower division, the Pulpit conglomerate, is named from the well-known Pulpit Rock at the mouth of Echo Canyon (Pl. 1; Pl. 11, fig. 1). The Saw Mill conglomerate in Saw Mill Canyon consists of 400 feet of massive, red, cliff-making conglomerate beds at the base, 500 feet of lighter-tan and gray conglomerate, grit, sandstone, and shale beds in the middle, and about 300 feet of brownish-tan conglomerate at the top. These beds, when traced up Echo Canyon, beyond the limits of the map, come down to and below the canyon floor in a broad syncline. Upon reappearing, however, on the northeast side of the syncline no unconformity is seen at the expected horizon. It was concluded that the Saw Mill and Pulpit conglomerates are here conformable and compose one unit sequence. Continuing up Echo Canyon the crest of a broad anticline is passed over before Castle Rock is reached (beyond the limits of Pl. 1), and at least 1000 feet of beds is exposed in the anticline below the approximate position of the contact of the two conglomerates. From Castle Rock to Wasatch Station the beds dip gently upstream so that the stratigraphic position of the beds at the roadside at Wasatch Station (from which Hayden took the name Wasatch group) is probably slightly

below the unconformity in Saw Mill Canyon. The beds change progressively from the extremely coarse conglomerate in the lower part of Echo Canyon to shale and sandstone at Wasatch station. Red coloring is irregular horizontally and vertically, fading and alternating with gray, tan, and yellow.

One thousand feet of the Pulpit conglomerate is exposed at Pulpit Rock, but an additional 1000 feet probably is present in the trough of the Stevenson Canyon syncline (Pls. 1, 4; cross sections D′–D″, E–E′).

Pebble counts were made above and below the unconformity in Echo Canyon. About 100 feet above the highway in the Pulpit conglomerate two thirds of the boulders and pebbles are Weber quartzite, and one third is limestone from the Brazer, Park City, and Twin Creek. Higher stratigraphically and just below the unconformity two thirds of the boulders again are quartzite, but here they have been equally derived from the Weber and Tintic. The Tintic quartzite boulders are similar to the Tintic quartzite that crops out in Hardscrabble Canyon, 20 miles to the west. Of the other one third most are sandstone and limestone boulders from the Three Forks, Brazer, and Park City. Nugget sandstone was identified, and a few black chert pebbles were seen. Some of the larger quartzite boulders that ranged up to 3 feet in length may be Proterozoic.

The boulders in the conglomerate above the unconformity are nine tenths purple and pink laminated quartzites and quartzitic conglomerates. These may be either Cambrian or Proterozoic, but the lithology is more characteristic of the Proterozoic. A few unique boulders from the Proterozoic of the northern Wasatch were found such as green quartzite, laminated phyllite, and red quartzite with quartz veins. These latter types of rock do not exist in the Uinta Mountains, as far as the writer knows, and suggest a source in the western highland. This suggestion was definitely confirmed when a number of boulders of schist, gneiss, and granite were found which are identical with the rocks of the Farmington Canyon complex.

The remaining one tenth of the boulders are of Paleozoic and Mesozoic derivation. In the narrows of the Weber Valley just above the cement plant at Devil's Slide the boulders are subangular and composed over one half of limestone from the Twin Creek, Thaynes, Park City, Brazer, and Madison formations. The other half consists of quartzite boulders mostly from the Weber and sparingly from the Cambrian. Since these are the rocks that make up the islandlike mountain mass to the west which the conglomerate overlaps, the local mountain here was the chief source of the boulders.

Beds of low-rank bituminous coal are present in the Almy Canyon conglomerate in Lost Creek Canyon and its tributary, Toone Canyon. A few small mines have been opened at both localities. The coal bed in Toone Canyon is 6 feet thick, contains much "bone," and is overlain by a blue clay containing fossil plants. Above and below the coal and clay is coarse conglomerate. The stratigraphic position of the coal is probably about that of the lower part of the Saw Mill conglomerate. The coal bed at the junction of Hell Canyon and Lost Creek is only 75 feet above an unconformity separating truncated Twin Creek shales from the Almy conglomerate. The coal bed is 4½ feet thick with a 3-inch bone bed in the middle. A very coarse basal conglomerate underlies the coal, and it is overlain by conglomerate, sandstone,

and shale beds. Both beds of coal are probably of short horizontal extent and appear to be at different stratigraphic positions in the conglomerate.

Fowkes (?) formation.—Above the Echo Canyon conglomerate is the Fowkes (?) formation consisting of 800 feet of friable gray grit and sandstone and poorly indurated shale. Some of the shale is black and paper-thin. Fossil plants are abundant in a layer about 300 feet from the bottom, but the enclosing rock is so fragile that good specimens are difficult to preserve. The rocks are in large part of volcanic origin. Although highly altered the material was once largely volcanic ash probably of rhyodacitic composition, admixed with other sedimentary material and deposited in a body of water. The Fowkes (?) formation is best exposed downstream from the East Canyon Reservoir where it cuts across East Canyon. The formation appears to lens out southward and extends northward to Devil's Slide where it has been removed by erosion at the margin of the small Paleozoic and Mesozoic highland of Wasatch time.

A light-gray formation of rhyolitic and latitic ash and thin white sandstone beds overlies the Almy formation conformably in southeastern Wyoming. It is called the Fowkes formation (Veatch, 1907, p. 91–92). The similar lithology and stratigraphic position of the Fowkes in southeastern Wyoming and the East Canyon suggest a similar age. The Fowkes formation has recently been classified as upper Paleocene, and the Almy as middle Paleocene (Wood *et al.*, 1941, Pl. 1, p. 1).

EOCENE SERIES

Knight (?) formation.—Overlying the light-gray Fowkes (?) formation is a deep-red series of conglomerate, sandstone, and shale. It has been folded into a large syncline in which Morgan Valley and the lower part of East Canyon have been eroded, and it crops out on both sides of these valleys and extends high up the east slope of the Wasatch Range, even to the top in the headwaters of Hardscrabble and City Creek canyons (Pl. 1, D–D′). It is probably thickest along a section extending eastward from Porterville in East Canyon where a projection of surface dips indicates 5000 feet of beds.

The pebbles and boulders in the Knight formation in some places are dominantly from the Tintic quartzite and in others from the Paleozoic limestones. At the mouth of Hardscrabble Canyon several gneisses from the Farmington Canyon complex were noted, and one phyllitic arkosite from the Proterozoic sequence was identified. Other quartzite boulders may be from the Proterozoic also.

The conglomerate, sandstone, and shale beds of Morgan Valley occupy the stratigraphic position of the Knight formation (lower Eocene) of southeastern Wyoming and have the same characteristic red color but differ in being more conglomeratic and having no limestone. The Knight in southeastern Wyoming is but little folded, with maximum dips of 5°. In north-central Utah it is considerably folded with dips exceeding 25°. The Knight overlies the Fowkes and Almy unconformably according to Veatch. The lens shape of the Fowkes (?) in East Canyon suggests an unconformity at the base, but such a structure is not apparent. G. B. Richardson, P. V. Roundy, and G. R. Mansfield (Mansfield, 1927, p. 108) re-examined parts of the Wasatch group described by Veatch and came "to the view that the Fowkes

formation may in reality be a lens, and that the postulate of unconformity between the Knight and Almy formations which closely resemble each other lithologically, may not be compulsory." It has already been noted that in the north-central Wasatch the Fowkes appears to be lens-shaped, and the overlying Knight and underlying Almy are not separated by an angular unconformity.

OLIGOCENE SERIES

Norwood tuff (new name).—A formation chiefly of volcanic derivation is inlaid in the Knight (?) formation of lower East Canyon and Morgan Valley and in the Henefer formation of Little East Canyon. It was observed by Hayden (1869, p. 92) who gave it the name Salt Lake group. His description follows:

> "In the valley of Weber River, from Morgan City to Devil's Gate, there is a thickness of 1,000 to 1,200 feet of sands, sandstones, and marls of a light color for the most part, which I regard as of upper Tertiary age. These newer beds must have not only occupied this expansion of the Weber Valley, but also all of Salt Lake Valley, for remnants of it are seen all among the margins of the mountains inclosing Salt Lake Valley. . . . I found this series of beds so widely extended and so largely developed in Weber Valley and Salt Lake Valley that I regard it as worthy of a distinct name and in consequence have called it the Salt Lake group."

Peale (1879, p. 588, 649) extended the name to southeastern Wyoming, and Mansfield (1927, p. 110) modified it from Salt Lake group to Salt Lake formation because he believed a single diastrophic and stratigraphic unit was represented in southeastern Idaho, that the beds were chiefly of fluviatile origin and not lacustrine as Peale has inferred, and that the word formation was the proper designation for beds of varied character. In retaining the name Salt Lake he regarded the southeastern Idaho deposits as equivalent to the light-colored Morgan Valley beds of Hayden and assigned tentatively a Pliocene age to them. All fossils found were gastropods and pelecypods from which either early or late Tertiary age could be inferred (Mansfield, 1927, p. 111, 112). The beds were horizontal, inclined, and vertical in Idaho.

Vertebrate remains were discovered in the formation south of Morgan by John Clark and the writer in August 1940. The repository is located in a fresh road cut on the main East Canyon highway just below Norwood Canyon. An extract of a letter from Clark containing his identifications states:

> "The Titanothere is *Allops marshi*, typical of the lower Oligocene or Chadron formation of South Dakota and Nebraska. The artiodactyl is *Leptomeryx*, species indeterminable; this genus occurs throughout the Oligocene and even into the Gering formation of basal Miocene age. However, it is characteristically lower and middle Oligocene. The two together make the age of the Norwood tuff unquestionably lower Oligocene, and almost certainly equivalent to part of the Chadron formation of the Great Plains. It is, therefore, probably a little younger than the poorly known basal Oligocene Duchesne River formation of the Uinta Basin, from which it differs completely in lithology."

The beds in which the bones occur are made up of quartz sand with considerable white powder, numerous gray quartzite pebble lenses, and poorly consolidated tuff. The tuff is pale greenish tan at the north end of the road cut but changes bodily—not in single beds—to red and pink in the south end. The deposit is much channeled; and the bones probably were buried in the fill of a large eddy scour of a river.

The formation as a whole in the back valleys of the north-central Wasatch is composed dominantly of light-colored tuff with lenses of volcanic conglomerate that

become more abundant toward the south. Some of the tuff layers are of uniform thickness and wide extent. A fairly compact greenish tuff that forms a ledge over the low divide from Ogden Valley to Morgan Valley is an example. Microscopic examination shows it to be a very fresh rhyodacite tuff, with none of the common sedimentary clay minerals. This particular bed probably represents a volcanic dust fall in a temporary lake several miles across. The deposit as a whole is not well exposed, but numerous outcrops show channeling, and the formation is believed, therefore, to be in larger part fluviatile than lacustrine.

Under the microscope the pebble and boulder specimens all consist of a silicic glassy groundmass with embedded phenocrysts and microliths of plagioclase ranging from calcic andesine to calcic oligoclase, augite, hypersthene, hornblende, and biotite. The mineralogical composition of the tuffs is very similar to the pebbles, and there can be little doubt that the tuffs were derived from the same material.

Petrographic study also indicates that the tuffs of East Canyon and Morgan Valley are very similar if not identical with the andesite tuffs and breccias of the Park City district (Boutwell, 1912, p. 70–74). Although the Park City and Kamas Prairie volcanic field is not connected with the East Canyon and Morgan Valley tuffs, the short distance and the scattered erosional remnants between, and the similar lithology indicate a common volcanic parent. The fact that some of the tuffs are direct sediments of volcanic eruptions, and not secondarily reworked volcanic deposits, suggests a contemporaneous origin with the Park City volcanic deposits, a large part of which is also tuff. The breccias, agglomerates, and flows in the Park City area indicate that the approximate center of volcanic activity was there.

Hyrum Schneider (personal communication) believes that all the Park City volcanic rocks may not be equivalent in age to the Norwood tuff, even though all those described by Boutwell are similar.

The tuffs are separated by a profound unconformity from the underlying beds. They rest in a synclinal valley (Morgan Valley) on beds of the Wasatch group and in an anticlinal valley (Little East Canyon) on the Henefer formation. Folding and erosion followed deposition of the Wasatch beds and by the end of Eocene time wide deep valleys existed. Volcanic eruptions flooded the back valleys of the central Wasatch from Heber Valley to Ogden Valley, a distance of 70 miles, at the beginning of Oligocene time, and sediments accumulated in them which may have been 2000 feet deep in places. The pre-Norwood surface and related diastrophism are somewhat more complicated than the above brief indicates and are considered more thoroughly under the heading of physiography.

The tuff represents a distinct and important lithologic unit, and it is one of the few deposits in this part of the section that has yielded vertebrate fossils by which a definite age could be established. It is here proposed to call the beds the Norwood tuff from Norwood Canyon where they are well exposed, and near the mouth of which the titanothere bones were found. The Norwood tuff is recognized as a division under Hayden's Salt Lake group. When the intermontane valleys west of the Wasatch become better known stratigraphically, the Salt Lake group may be found to include several formations, perhaps separated in age by considerable parts of the Tertiary. This possibility emphasizes the necessity of a new formational name.

PLEISTOCENE AND RECENT DEPOSITS

Lake Bonneville sediments.—Delta, shore-line, and bottom deposits at the Bonneville and Provo levels of Lake Bonneville are shown on Plate 1. They have been adequately discussed by Gilbert (1890, p. 163–164) and seem most appropriately considered under the subject of physiography in this report.

Valley alluvium.—The post-Bonneville history has been chiefly one of erosion, but flood plains and small alluvial fans are common. Talus slopes exist around the high peaks. As Lake Bonneville shrank the desiccated remnant, Great Salt Lake, continued to receive sediments and thus contains an interesting and extensive sedimentary record. Only the topmost layer has been studied (Eardley, 1938).

STRUCTURE

GENERAL FEATURES

The following description of the general structural features of the north-central Wasatch may be made clearer if the reader will turn to the geologic maps (Pls. 1, 2) and the tectonic map (Pl. 3) before proceeding with the text.

A mass of pre-Cambrian crystalline rock in the western part of the area forms a nucleus around which compressional structures in the younger overlapping stratified rocks are shaped. At the northern and northeastern margin of the crystalline mass are the thrusts of the Ogden Canyon, Ben Lomond, and Durst Mountain areas. Around the south and southeast margin are the folds of the central Wasatch. Extending eastward from the southern limit of the area is the Uinta uplift. It forms an L with the north-south trending crystalline mass. In the angle of the L are the Tertiary folds of the Henefer and Coalville area. These folds appear to be a continuation of the westward-veering fold and thrust structures of the Idaho-Wyoming arc. The Tertiary folds are discordant in trend to the Uinta uplift but in part are continuations of the folds of the central Wasatch. The folds of the Idaho-Wyoming arc in the northern part of the area trend parallel with those of the central Wasatch and meet the Ogden Canyon thrusts approximately at right angles. This area is one of complex superposition of compressional structures and reveals a growth in at least eight stages of deformation. The ages of the stages of deformation are revealed in the relations of the Cretaceous and Tertiary rocks in the Henefer and Coalville area.

During the crustal movements certain areas were subjected to erosion and others to deposition. The deposits were in turn deformed, eroded, and their waste products laid down in adjacent subsiding areas. Large parts of complex older structures are covered by less intensely deformed younger strata. The structures in the younger beds both parallel and cut across the trend of the covered structures. If parallel in trend they may be offset or even antithetical—*i.e.*, an anticline over a syncline.

THRUSTS

Ogden Canyon thrust area.—The thrusts of Ogden Canyon are well known from the work of Blackwelder (1910a, p. 517–542) 30 years ago. His reconnaissance map shows the major structures. The details have been traced by the writer on photographs (Pls. 4, 6) and incorporated in Plate 1.

Three thrust sheets of major proportions make up the range immediately north

and south of Ogden Canyon. The lower two have an imbricate arrangement, dip east-northeastward, and are sharply folded transverse to their general strike. The highest rests across the beveled edges of the lower two and has a gentle eastward dip. The position and relation of the three thrust surfaces are shown in the block diagram of Plate 1. The structure along the north wall of Ogden Canyon is illustrated in the cross section immediately above the block diagram. Two other cross sections, A–A′ and B–B′, of the Ogden Canyon area appear in the same plate.

The lowest fault, first crossed in ascending Ogden Canyon, will be called the Taylor thrust because its trace parallels the lower course of Taylor Canyon, about 1½ miles south of Ogden Canyon. The next fault up the canyon is designated the Ogden thrust, a name first given by Blackwelder (1910a, Fig. 7, p. 537). The Taylor thrust can be traced up the north canyon face (Pl. 4) and westward down the mountain front to the Bonneville beach, where the Tintic quartzite, in sedimentary contact with the Farmington Canyon complex, is obliquely cut off with little drag. The thrust here dips 50° N. The base of the Tintic marks the Ogden thrust as far as it can be traced northwestward from the canyon bottom. Blackwelder shows an east-west transverse fault where the Taylor thrust comes to the west foot of the range. By means of it the quartzites are so displaced that the Taylor thrust seems to continue as the Ogden thrust. The aerial photographs on the other hand show no break in continuity of the quartzites above each thrust, and field observations correspond with those made on the photographs. It is concluded that no transverse fault exists and that the northward-dipping Taylor thrust at the mountain front is the fault that Blackwelder mapped as a transverse fault.

The Taylor and Ogden thrusts may also be traced southward from the canyon bottom on the aerial photographs. The faults and the contacts of the quartzite and certain beds of limestone are revealed in exquisite detail. From Ogden Peak or from the canyon bottom the quartzite extending northwestward from Ogden Peak appeared to be continuous with the first quartzite in Ogden Canyon. The photographs clearly show that it is cut out for a short distance, and limestone beds of entirely different strike and dip occupy the gap. The Taylor thrust, as on the north side of the canyon, continues westward to the mountain front but dips southward, thus defining a pronounced eastward-pitching anticline in the thrust surface. The Ogden thrust parallels approximately the Taylor thrust and exhibits the same folded nature. A transverse fault was mapped by Blackwelder along the trace of the westward-veering Taylor thrust, and he named it the Huntsville fault. Without the aid of the aerial photographs the writer probably would not have recognized the folded thrust structure and would have followed Blackwelder in mapping a transverse fault. The results of detailed mapping lead to the following interpretations: (1) The uppermost quartzite in Ogden Canyon connects underground with the quartzite through Ogden Peak; (2) the middle quartzite in Ogden Canyon connects with, and is in the same thrust sheet as, the quartzite along the mountain front between Ogden Canyon and Taylor Canyon; (3) the lowermost quartzite in Ogden Canyon crops out again near the mountain front between Taylor Canyon and Waterfalls Canyon. Thus, instead of a transverse fault, the structure is one of transverse folds in the thrust sheets, as illustrated in Plates 1 and 5.

The map also indicates that a cross-trending anticline and syncline were sliced by the Ogden and Taylor thrusts. The position of the cross folds in the Ogden thrust sheet is just south of the Hermitage of Ogden Canyon. The axes run approximately east-west and are offset southward in the underlying Taylor sheet (cross section B–B′, Pl. 1). Assuming that the cross folds of the thrust surfaces indicate that the thrust sheets were folded after the thrusting and not that the sheets developed originally with gigantic mullions, the pre-folded condition can be restored and the nature of the pre-thrusting cross folds determined. This was done, and as a result the anticline about disappeared and the syncline remained as the conspicuous fold between nearly horizontal strata. It is concluded that an early east-west-trending syncline existed approximately along the site of the present Ogden Canyon before the thrust faulting began.

Quartzites, phyllites, and argillites of Proterozoic (?) age rest on the truncated edges of all the rocks involved in the Taylor and Ogden sheets. Blackwelder recognized this relationship, interpreted it as a thrust, and named it the Willard overthrust. The low-grade metamorphosed strata of the Willard thrust sheet break down into fairly smooth and light-colored slopes. Their base, and hence the trace of the Willard thrust, is easily followed. In addition to the distribution that Blackwelder showed, the floor of Wheeler basin is underlain by the Proterozoic rocks of the Willard thrust sheet. What was once a sheet over 15,000 feet thick has been eroded to a very thin capping or entirely removed. Where the beds still remain they are badly contorted. The Willard thrust is of much greater stratigraphic and horizontal displacement than the Taylor or Ogden thrusts.

The thrust faults in Ogden Canyon, even though conspicuous and well exposed, are hardly as impressive to look at as the Z-fold up canyon from The Hermitage (Pl. 1; Pl. 6, fig. 1). The view of the Z-fold on the north side of the canyon just below Pine View Dam is particularly striking, and is pictured by Blackwelder (1910a, p. 538, Pl. 40, fig. 1). The axial planes of the fold are nearly horizontal, and the axes trend approximately N. 40° W. The axial plane of the upper fold of the Z is only a few feet above the canyon bottom at The Hermitage but gains altitude upstream so that at a point a short distance below the Pine View Dam it is 300 to 400 feet above the river. It is generally so near the canyon bottom that talus conceals it, and its continuity from The Hermitage to Pine View Dam is not generally observed. The Z-fold is much better exposed on the north side of the canyon than on the south, although an impressive view of overturned recumbent beds may be had on the south side just a short distance up the canyon and across from The Hermitage near the river level. Beds from the lowest Cambrian limestones to, and including, the Brazer formation are involved.

The Z-fold may appear to result from drag of the upturned beds under the Willard thrust sheet. It appears that Blackwelder (1910a, p. 528, Pl. 40, fig. 2) first visualized the thrusting as toward the southwest because of the northeast dip of the thrust planes. Schneider informs the writer that he and Blackwelder later visited the area together and were so impressed by the Z-fold that they were prone to think at the time it was the result of drag under the Willard thrust sheet moving approximately east.

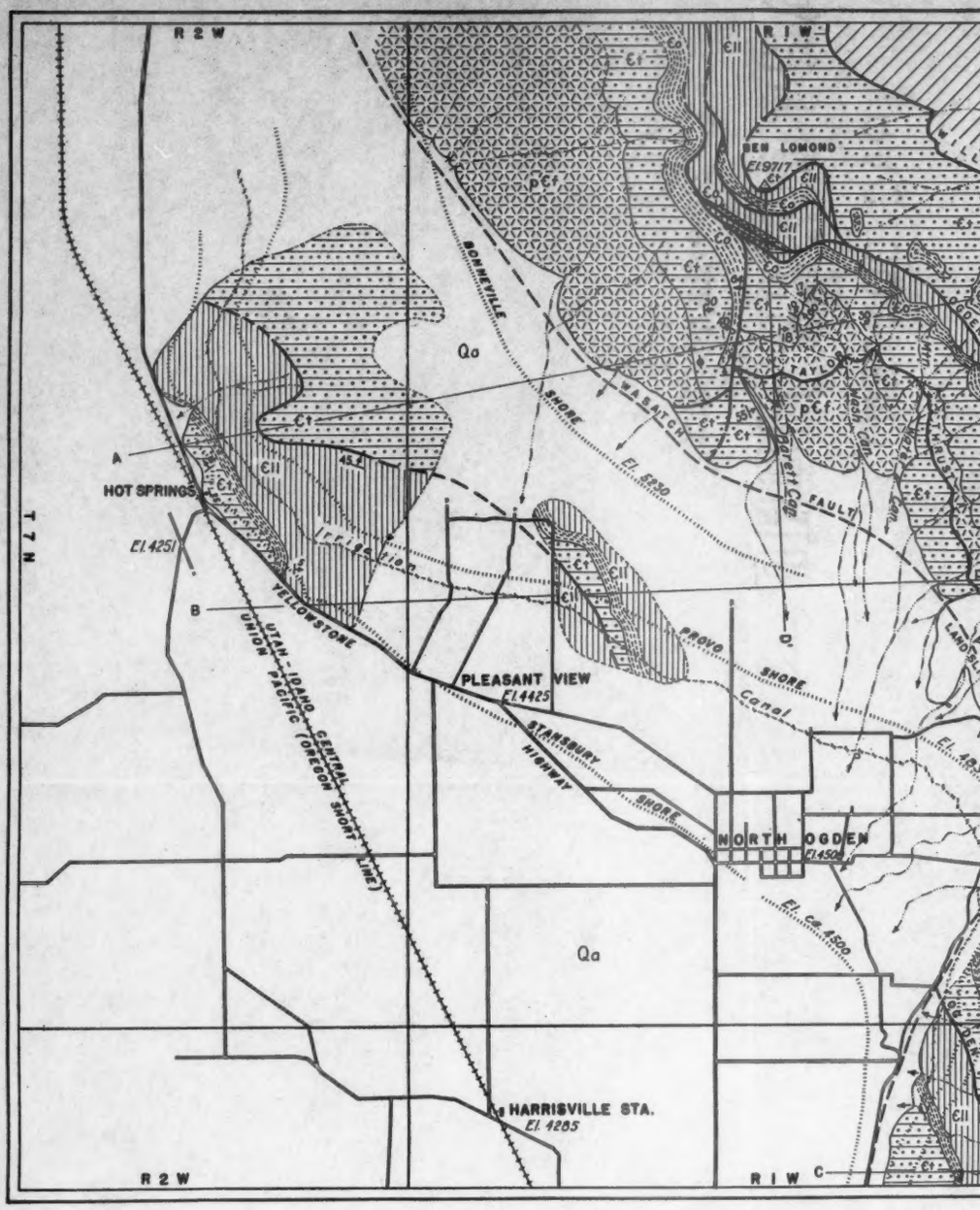
The writer believes that if the Z-fold were the result of drag beneath a thrust sheet moving east the upper horizontal of the Z would probably not exist because it stands directly opposed to the overriding sheet. On the other hand, the Z-fold cannot logically be accounted for by movement of the overriding sheet to the west, but if the fold already existed it would not oppose movement in that direction.

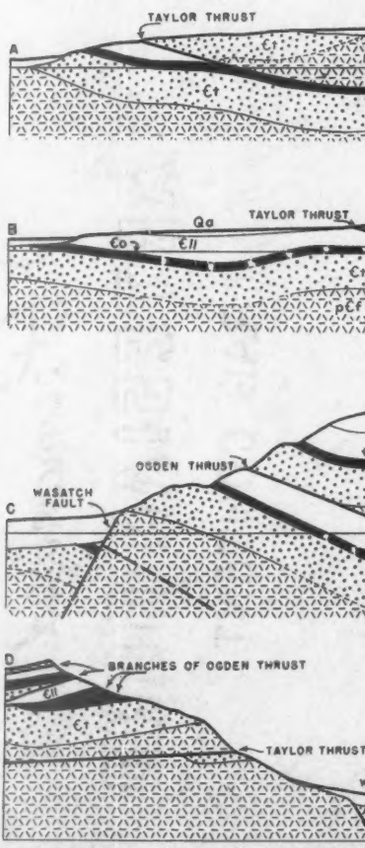
The writer (1939, p. 1286) compiled a tectonic map of the Wasatch-Great Basin region and postulated a small, fairly resistant shield, the northern Utah highland, as existing during the Laramide revolution. About the margins of this shield the sedimentary rocks have buckled and sheared. The Taylor and Ogden thrust sheets overrode the northeastern margin of the highland, according to this theory, and moved from east-northeast to west-southwest. Problems of this hypothesis are discussed later under the heading, Relation of local structures to regional. If the Z-fold is related to the Taylor and Ogden thrusting it is probably an incipient shear which formed at the same time as, and complimentary to, the Ogden thrust. Folding to the stage of a sharp-crested Z occurred, with the axial planes dipping just opposite and about the same amount as the incipient Ogden thrust. The Z-fold did not rupture under further compression, but the incipient Ogden thrust broke and partially relieved the strain (Pl. 1, section along north face of Ogden Canyon). As the Ogden thrust was displaced, the Z-fold was rotated clockwise until its axial planes were nearly horizontal, and now nearly parallel with the gradient of Ogden Canyon. Erosion then truncated the strata to the top of the Z-fold before the Willard thrust sheet rode ahead.

Ben Lomond thrust area.—Ben Lomond is the high peak northeast of Ogden and east of the Pleasant View salient. The area designated by the name extends southward from the commanding peak to Eden Pass where it joins the Ogden Canyon thrust area. The rocks of the area are complexly faulted and are shown on a separate geologic map (Pl. 2) of larger scale than the map of the main central Wasatch (Pl. 1).

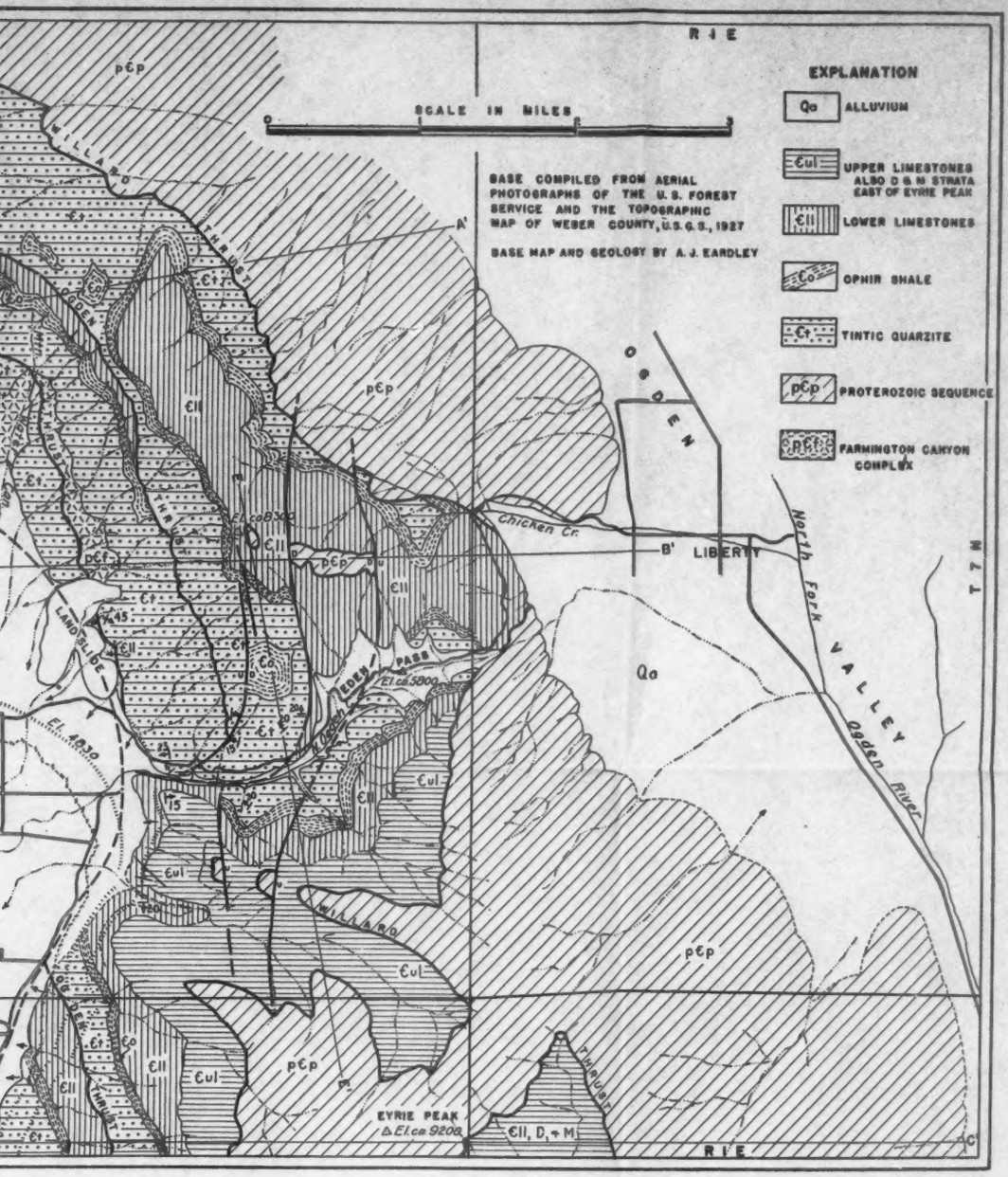
Mapping of the main Wasatch Range north of Eden Pass revealed three thrust faults in approximately the same relation to each other as in Ogden Canyon. The Willard thrust continues through and is thus definitely established, but the two lower thrusts when projected did not meet those of Ogden Canyon. After a cross fold through Eden Pass was worked out it seemed probable that the Taylor and Ogden thrusts from the mountain front just north of Ogden Canyon curve northward and then eastward and connect with the two lower thrusts of Ben Lomond. Cross sections demonstrate the possibility, in fact probability, of this interpretation (Pl. 5), and in the regional fault map (Fig. 3) the connections are so drawn (T. 7 N.). The cross fold will be described later.

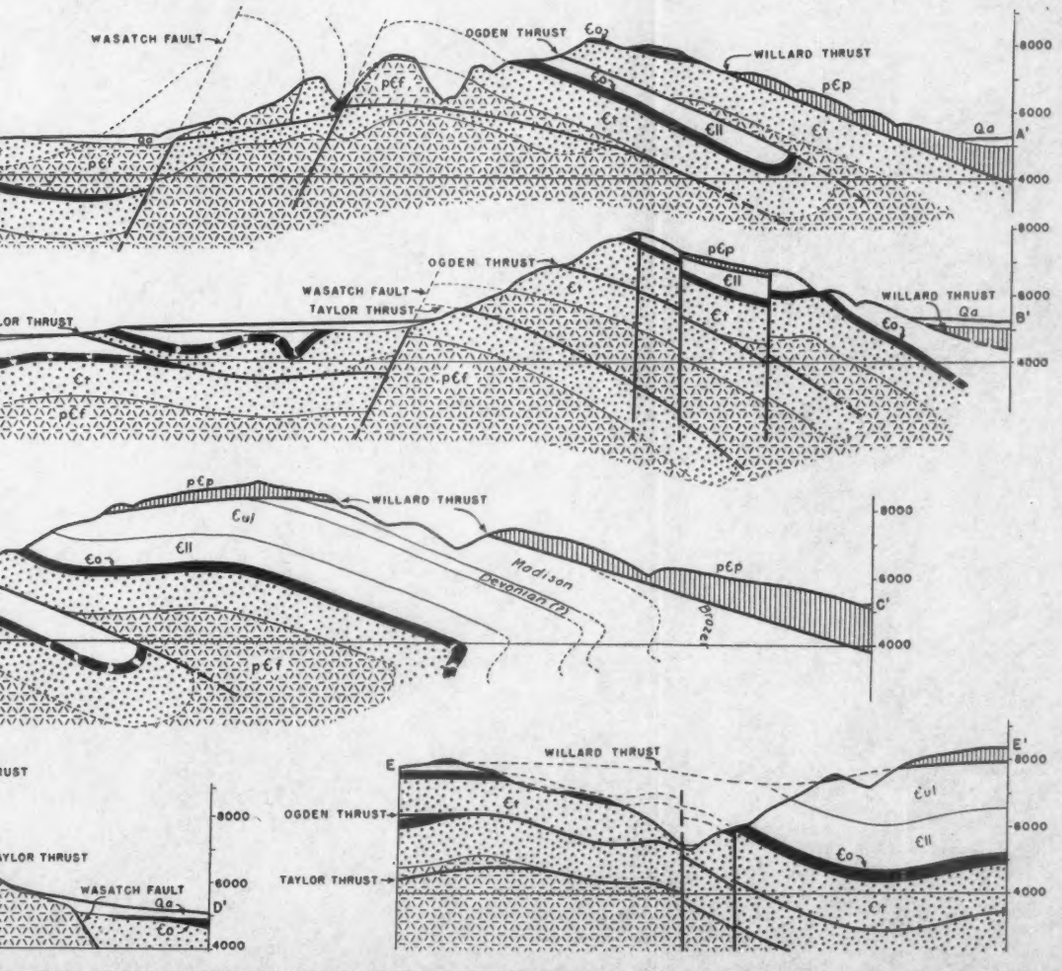
The Taylor thrust thus recognized emerges from under the alluvium at a landslide on the mountain just north of the road over Eden Pass, and it may be followed northward at a rather low elevation on the mountain front to Water Canyon. From there it extends westward to and across Barett Canyon. It dips about 30° into the mountain south of Water Canyon but westward from there to Barett Canyon it is nearly flat (Pl. 2, A–A′). In Barett Canyon it is cut by a north-south high-angle fault which drops the west side about 200 feet. The thrust dips gently westward from this point until it disappears under the alluvium.





GEOLOGIC MAP AND CROSS SECTIONS OF THE BEN





HE BEN LOMOND AREA, WASATCH MOUNTAINS, UTAH

The Cambrian formations are repeated in the Pleasant View salient, and the irregular contact between the two sections indicates a gently dipping fault plane. The abundant deposits of Lake Bonneville conceal much of the bedrock, and the interpretation of a thrust fault relation is, therefore, not certain. To assume that it is a low-angle thrust helps considerably in constructing a cross section, and in the interpretation of section A–A′ it is shown as the Taylor thrust. The alternate possibility of repetition by a normal fault was tested in cross section, but it seemed impossible to get the Cambrian beds in the salient, below the Farmington Canyon complex in the mountain, without a fault of prohibitive throw. There is also no scarp to mark the trace of a normal fault if it is of Basin and Range age.

The cross section as presented shows a gentle north-south anticline and syncline in the thrust surface. A gentle anticline, the Eyrie swell, was recognized in the Willard thrust surface to the south, and it appears that the Eyrie swell extends to the anticline in the Taylor thrust.

The Ogden thrust emerges from the cross fold in the lower part of Eden Pass Canyon, where the Tintic quartzite is repeated. Almost two full stratigraphic sections of quartzite are present there. As the fault gains elevation northward on the mountain front the two quartzites are separated first by the Ophir shale and then by the shale and additional beds of Cambrian limestone. The fault reaches almost to the crest of the range near the head of Water Canyon, where it splits into four branches. The branches bound slices of the Ogden thrust sheet which make up the mountain mass immediately under Ben Lomond (Pl. 2, D–D′). Since the branches may be observed along the trace of the Ogden thrust, it is inferred that they join the main fault in their dip direction also. The Ogden thrust has not been mapped north of Ben Lomond, but, as the trace of the Willard thrust lies diagonally across the crest of the range in this area and passes under the alluvium just north of Willard Canyon, probably the trace of the Ogden thrust terminates at an acute angle with the margin of the Willard thrust sheet. If so, this represents another place where the Ogden sheet is truncated by the Willard (Fig. 3, T. 8 N., R. 1 W.; Pl. 5).

The Willard thrust was mapped rather accurately by Blackwelder (1910, p. 536) in the Ben Lomond area, and the present map adds only a few interesting details. A piece of the thrust sheet lies separated from the main mass in a small graben across a ridge crest just north of Eden Pass. A small klippe reposes on the Ogden thrust sheet a short distance to the west. Both outliers are crossed on the trail from Eden Pass to Ben Lomond.

A long finger of the Willard thrust sheet that extends westward from Eyrie Peak has been broken in two places and dropped 100 feet or so on the west by the same group of high-angle faults that created the graben.

A tear fault through the Ogden and Taylor thrust sheets could not exist down the canyon from Eden Pass because the formations are not cut and offset. Still the traces of the thrust faults seem to end at the canyon. A cross flexure with downfolding on the south would cause the Ogden thrust to curve sharply down the bottom of the Eden Pass Canyon. With a later high-angle fault also cutting down part of the canyon and the dropping of the south side still more, the cause for the termination of the thrust at the canyon bottom is still better explained. The cross flexure

as visualized is depicted in Plate 5, and the imagined positions of the traces of the Ogden and Taylor thrusts, thus flexed and pitching, are dashed on Figure 3.

Durst Mountain thrust area.—Durst Mountain is the highest point of a pre-Cambrian, Paleozoic, and Mesozoic rock island surrounded by Cenozoic sediments (Pl. 1). The mountain mass lies east of Morgan Valley and represents the beginning of bifurcation northward of the Wasatch Range. The Laramide structure is, for the most part, one of eastward-dipping beds including all formations from the Tintic quartzite to the Twin Creek (Pl. 1, Secs. CC′, DD′). Section DD′ reveals a pitching anticline and syncline at the town of Morgan. Farther up the canyon the beds flatten at the Weber quartzite exposures but then roll over again to a 65-degree dip. They are overlapped irregularly on the east by the Almy conglomerate which blankets successively the formations from the Twin Creek at Croyden to the Brazer at Cottonwood Creek.

The main feature of the structure at the western front of the mountain just east of Peterson is a thrust fault of Laramide age. It is mostly obscured by the Knight formation and the Norwood tuff. The thrust appears as a transverse fault in Spring Hollow and is the site of a small strike valley from Spring Hollow to Bohman Hollow where the low northward dip of the reverse fault and the excessive brecciation of the rocks are especially clear and indicative of thrusting. At Spring Hollow, however, the relation of the crystallines on the south side to the Cambrian strata on the north side is not altogether clear because the fault follows the bottom of the canyon. If the fault plane dips southward, the Cambrian beds will be cut off in similar manner as at Bohman Hollow—*viz.*, the strike of each fault is oblique to the strike of the beds, and each fault is a reverse fault. A logical interpretation of the relations apparently is that the Bohman Hollow thrust arches over and connects with the Spring Hollow fault to form a folded thrust surface with an eastward pitch. Such a thrust would be of the same nature and generation as the Taylor and Ogden thrusts in Ogden Canyon.

The Cambrian quartzite exposures north of Bohman Hollow cannot be accounted for by block faulting because a vertical displacement of several thousand feet would be required to depress the quartzite of Durst Mountain to the western exposures on Cottonwood Creek. As the correlation of the two quartzites seems certain, another thrust probably branches off the thrust in Bohman Hollow (Pl. 3). Such a fault is nowhere clearly exposed, but it may run northward under the Knight formation and skirt the quartzite on the north side of Cottonwood Creek, perhaps entirely within the pre-Cambrian crystalline rock.

At the mouth of South Fork of the Ogden River a small nose of Cambrian limestone crops out. The beds dip southeastward as if projecting under the great Proterozoic sequence in the mountain immediately on the east. This relation was interpreted by Blackwelder as due to the Huntsville fault, a transverse fault of Laramide age that he believed extends from the south side of Ogden Canyon eastward past Huntsville to the nose of Cambrian limestone. The detailed mapping of the Ogden Canyon thrust structure demonstrated the absence of such a fault, and plane-table work shows that such a transverse fault would have to curve incredibly to separate the Cambrian limestone from the Proterozoic strata. Since the

Proterozoic mass is the overriding sheet of the Willard thrust, and since the southward trace of the Willard thrust is concealed at the margin of the Tertiary deposits east of Ogden Peak and normally should run for many miles before dying out, the geology at the South Fork locality seems to reveal the location and the continuation of the thrust. If it extends from Ogden Canyon to the mouth of South Fork the great thrust surface must be gently downwarped along the axis of Morgan Valley and Ogden Valley (Pl. 1, sec. AA′; Pl. 3). Downwarping of the overlying Tertiary deposits will be described later. The gentle but broad syncline in the thrust sheet is labeled the Huntsville sag on Plate 3. The trace of the Willard thrust is believed to extend eastward under the Norwood tuff between Morgan Valley and Ogden Valley, northeastward under Ogden Valley, and back around the pitching end of a small anticline in the thrust surface (sec. AA′) and through the saddle occupied by Tertiary conglomerate separating the nose of Cambrian limestone from the Proterozoic strata.

The alignment of the western margin of the Almy conglomerate east of Durst Mountain thus acquires significance. The trace of the Willard thrust should lie just east of the margin, buried beneath the conglomerates, and should extend toward the junction of Lost Creek and the Weber River. Immediately east of the junction the Henefer formation crops out and is folded into a pronounced asymmetrical anticline. At the junction a neck of Almy conglomerate separates the Henefer from the Twin Creek beds which are turned up on end and badly shattered (Pl. 1, sec. D′D″). As a thrust dies out stratigraphic displacement decreases until the youngest formations involved are in contact. This might be the situation near the Lost Creek and Weber junction, but the interpretation requires a dying out of the great Willard thrust in a very short distance. The problem will be discussed at length under the heading, Relation of local structures to regional.

Evolution of thrust structure.—A number of stages in the development of the thrust structure, especially of the Ogden Canyon area, can be recognized. By "stages" is not necessarily implied separate diastrophic impulses, for the growth process might have been continuous from beginning to end or continuous over certain groups of stages. However, the Taylor and Willard thrusts probably advanced over erosion surfaces, and these episodes of erosion suggest periods of cessation of compression.

The first stage of the structural development is the east-west folding indicated by the syncline sliced by the Taylor and Ogden thrusts. Since the Taylor thrust sheet on either side of the syncline rests upon the edges of the beds below after the manner of an erosion thrust (Pl. 1, D′–D″), some north-south folding must also have preceded the thrusting. This folding is postulated to have been simply upturning of beds at the margin of a small highland that lay to the west (Eardley, 1939, p. 1285–1289). Since the sequence of the two foldings, the east-west and the north-south, cannot be deciphered and also because very little is known about either, they are both assigned to the first stage. The thrusting of the Taylor sheet is considered the second stage. Beautifully displayed drag on the underlying beds on the south side of Ogden Canyon indicates westward movement. During this stage the Z-fold in Ogden Canyon is believed to have formed. Before the Taylor thrusting was completed, the Ogden thrust and the Durst thrust probably began to develop. In the

third stage the Ogden thrust sheet moved westward, it is believed, a few miles in the same direction as the Taylor thrust sheet and in so doing rotated the Z-fold axes clockwise to an almost horizontal position. The shear occurred mostly along the Ophir shale, but at the location of the cross syncline the Tintic quartzite, Ophir shale, and Cambrian limestone were cut obliquely. The Durst thrust sheet moved westward at the same time as, and in imbricate arrangement with, the Ogden sheet. All these events preceded the main cross folding which marked the next stage. The two thrust sheets in Ogden Canyon were strongly folded normal to their existing strike (Pl. 1, sec. B–B′ and block diagram; Pl. 5). A third thrust sheet, the Willard, that rests on the truncated edges of the lower two, does not partake of their sharp cross folding, and hence the cross folding predates the third stage of thrusting. Evidently, the main cross folding was followed by extensive erosion, which prepared a fairly smooth surface that the Willard thrust sheet later moved over.[3] The crustal shortening effected by the Willard thrust was probably much greater than that of the Taylor and Ogden thrusts together. The Willard sheet appears to have moved ahead very rapidly because no erosional debris from the front of the thrust sheet is known to have been overridden. The third episode of cross folding followed the Willard thrusting, and the Willard sheet was gently arched along Ogden Canyon directly over the more pronounced anticline of the two thrust sheets below. The anticline may be traced eastward to and along the course of the Middle Fork of Ogden River in the Proterozoic quartzites, and from there northward into the Bear River Range. The Willard thrusting is denoted as the fifth stage of deformation, and the third episode of cross folding as the sixth stage.

The thrust structure in the north-central Wasatch Range was completed by the sixth stage, except for gentle north-south folding that gave rise to the Eyrie swell and Huntsville sag. All the stages of thrusting and cross folding preceded Almy deposition. The several folds and unconformities in the Cretaceous and Tertiary formations east of Ogden Canyon must be described before the evidence for the dating of the thrusts can be given, and before the concluding stages of the Laramide revolution in the north-central Wasatch are added.

FOLDS

Coalville anticline.—The Coalville anticline at the town of Coalville crosses the Weber River and the lower course of Chalk Creek. About 10,000 feet of Frontier beds, 600 feet of Aspen beds, and 800 feet of Kelvin (?) beds are exposed in the anticline, and unconformable overlaps by the Henefer formation, the Pulpit conglomerate, and the Saw Mill conglomerate effectively circumscribe the Cretaceous and Jurassic rocks. The area was described by Wegemann (1915, p. 161–184), and a geologic map with overprinted structure contours is included in his report. The geologic map is difficult to interpret because the pre-Henefer succession was mapped

[3] Instead of an erosion thrust the Willard could conceivably be a subsequent shear thrust, as defined by Billings (1933, p. 145), and typified by the Bannock overthrust. The underlying beds of the Willard thrust are not folded like those of the Bannock, and pre-Cambrian rocks are involved in the Willard and not in the Bannock. Because of these differences and also because it is difficult to explain shearing along such a flat and extensive plane as the Willard, the writer believes the erosion thrust concept is the more logical. Positive evidence for either view is lacking, however.

as one unit, and the overlying Henefer formation, the Pulpit conglomerate, and the Saw Mill conglomerates were treated as another unit. On Plate 1 the writer has differentiated the formations just mentioned as well as the ridge-making sandstones in the Frontier. The remnant of the Almy conglomerate near the center of the structure lies in a pre-conglomerate depression. The axis of the anticline trends about N. 30° E. where exposed and pitches northeastward. The structure is decidedly asymmetrical; the southeast limb is slightly overturned in one place, but the northwest limb dips only 11°–24°. A prominent nose extends northwestward from the axis, and the open valley of Chalk Creek below the narrows is eroded in this nose. Most of the southwest extension of the anticline is covered, but a slight tendency toward closure in that direction is indicated by the bend of the outcrops toward the axis in Lewis Canyon.

As the Henefer formation overlies the Frontier unconformably (Pl. 1, sec. F–F′), folding and erosion of the Frontier beds preceded Henefer deposition. More folding and erosion occurred after the Henefer was laid down because these beds are overlain unconformably by the Almy conglomerate. The Henefer formation had been entirely eroded, and Frontier sandstones and shales approximately 10,000 feet below the top beds were exposed in part of the anticline before the Almy was deposited. However, most of the erosion in the Frontier preceded Henefer deposition.

Stevenson Canyon syncline.—The town of Echo lies on the axis of a syncline in the Pulpit conglomerate, and Stevenson Canyon which extends southwestward from Echo has been eroded along the trough. Echo Canyon at its mouth is in the trough of the syncline (Pl. 1, sec. D′–D″) but upstream cuts into the southeast flank of the syncline. However, the valley at this place is in the flat-lying beds of the Saw Mill conglomerate that rest unconformably on the Pulpit (sec. E–E′). The syncline is one of gentle dips and southwest of Echo is almost symmetrical.

Henefer anticline.—Running obliquely across Weber Valley is an anticline in the Henefer formation. The anticline has been traced from Henefer southwestward to Little Emigration Canyon and northeastward to Brewer Canyon and, perhaps, to Toone Canyon, both tributaries of Lost Creek. Its axis is approximately parallel to that of the Stevenson Canyon syncline and the Coalville anticline. Like the Coalville anticline, the Henefer anticline is distinctly asymmetrical with the southeast limb almost vertical (Pl. 1, sec. D′–D″). Between Wolf Den Canyon and Brewer Canyon the axis pitches steeply to the northwest and then rises gently to Toone Canyon. The lower part of Toone Canyon is eroded through the gently arched Almy conglomerate, which rests on the truncated edges of the Twin Creek (?), Entrada (?), and Kelvin (?) formations. At the mouth of Toone Canyon a sharp syncline exposes lower Kelvin (?) beds. Upstream the red shales and sandstones of the Entrada crop out, and the first gray limestone ledge that comes down to the road is at the top of the Twin Creek. About 600 feet of beds comprising the top part of the Twin Creek formation is folded into an anticline which parallels the syncline just mentioned. Both trend parallel to Lost Creek and lie on a projection of the Henefer anticlinal axis. The northwest flank of the anticline in the Twin Creek dips 60°, and the southeast flank is slightly overturned. The axis pitches about 30° SW., emphasizing the saddle in the Henefer anticline.

The Twin Creek shales and limestones are exposed almost continuously from Toone Canyon up Lost Creek to the forks where the Twin Creek and Nugget contact is located. It would be desirable to know the relation of the Nugget and Twin Creek contact at the forks with the anticline and syncline in the Twin Creek, Entrada (?), and Kelvin (?) formations in Toone Canyon, but a careful charting of strikes and dips of the narrow belt of outcrop of Twin Creek connecting the two localities (all that is exposed beneath the Almy conglomerate) was not particularly revealing. The structure of the strata under the conglomerate is undoubtedly very complex because the Willard thrust is only about 6 miles to the west, and the Bannock thrust extends into this area from the north and probably converges here with the Willard thrust. Any interpretation of structure beneath the conglomerate, even though very logical, would only be one possibility in many and, therefore, does not seem worth presenting. A conclusion worth emphasizing, however, is that the sharp folds in Toone Canyon and the development of fracture and possibly flow cleavage parallel to the fold axes are definitely indicative of complex structure under the conglomerate.

Morgan Valley syncline.—In Morgan Valley a wide syncline, chiefly in the Knight formation, extends southward up East Canyon to its tributary, Sheep Canyon, thence up Sheep Canyon to the divide, and ends with the termination of the conglomerate of the Knight in the headwaters of Mountain Dell Creek. It continues northward across the divide into Ogden Valley where alluvium covers any Tertiary rocks that may exist and by which the continuation of the syncline could be recognized. Dips of 25° on the west flank and 16°–25° on the east are believed to indicate folding and not initial dip.

The conglomerate at one time covered Durst Mountain and the crest of the Wasatch Range south of Bountiful Peak, because remnants of the conglomerate still exist in those high areas.

The conglomerate beds are approximately parallel to the floor upon which they rest, and there is no indication of sharp local relief just before their deposition. On the contrary, much of the present height of Durst Mountain on the east of the Morgan Valley syncline and of the Wasatch crest on the west may well be due to gentle folding that followed accumulation of the Knight formation and that recurred after the Norwood tuff was deposited. A great valley system, however, was eroded in the conglomerate after its first gentle folding. In it the Norwood tuff was deposited in early Oligocene time. Near the west front of Durst Mountain the tuff locally dips 45° toward the mountain. This is not due to block faulting because the fault block was tilted only a few degrees. It is inferred that local buckling occurred within the syncline.

Truncated Paleozoic and Mesozoic beds of Durst Mountain make up the gently arched core of an anticline in the Wasatch beds which lies adjacent on the east to the Morgan syncline (Pl. 1, sec. D–D′–D″). The main Wasatch Range west of the Morgan syncline is also probably the core of a gently arched anticline in the Tertiary beds. The axis of the Eyrie swell extends southward, and the axis of an anticline in the Knight formation between the Salt Lake salient and Hardscrabble Canyon appears to be the continuation (Pls. 1, 3).

Folds of the Central Wasatch.—Two sharp synclines and an intervening anticline

extend eastward from the central Wasatch front. Schneider's (1925, p. 33, Fig. 3) cross section shows synclinal troughs approximately at the mouth of Emigration Canyon and along Parleys Canyon, with an anticline in the ridge between. The folds are part of a much larger syncline between the northern Utah highland and the Cottonwood uplift (Schneider, 1925, p. 33; Eardley, 1939, p. 1282, Pl. 1). The two synclines and the anticline pitch eastward. Their axes veer to the north and line up with the Henefer anticline. However, they comprise beds of Jurassic and Triassic age, and these are overlain unconformably by the Wasatch and Henefer beds in which the Henefer anticline is found. The relation of the structures will be discussed later. The east-west folding in the central Wasatch occurred sufficiently before Henefer time to allow fairly deep erosion, at least locally, before the Henefer beds were deposited.

HIGH-ANGLE FAULTS OF LARAMIDE AGE

A high-angle fault displaces the Taylor thrust in Barett Canyon on the slopes of Ben Lomond (Pl. 2). It strikes nearly north and curves westward at the approach to the Ogden thrust and apparently merges with it. A mullioned wall recently exposed by a rock fall on the east side of Barett Canyon reveals a 60-degree dip to the west and dip-slip movement with the west side down. The movement as measured by the displacement of the Taylor thrust is about 200 feet.

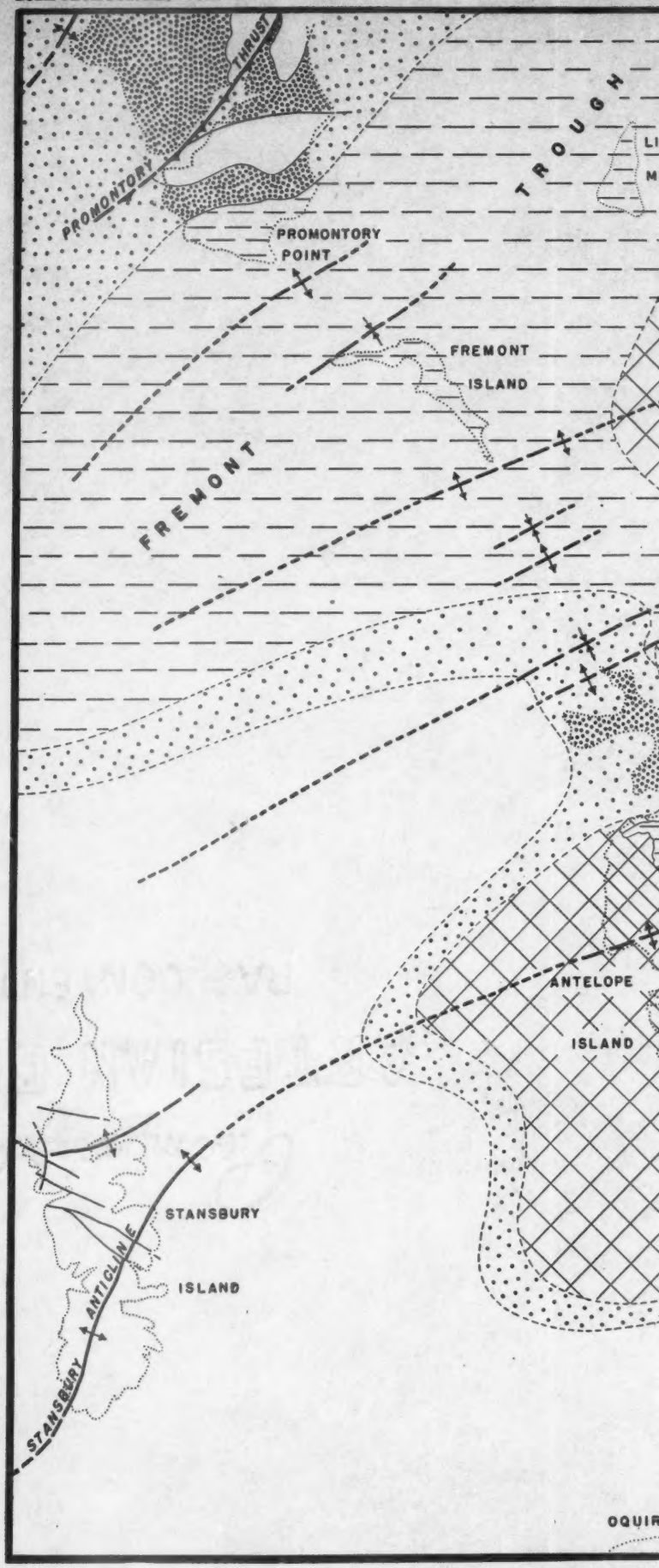
Five other high-angle faults probably related to the Barett Canyon fault have been noted at Eden Pass. Three of them cross the ridge north of the pass, and two cross the ridge south of the pass. Their position across the ridges is definite because remnants of the conspicuous Willard thrust sheet are cut off, but just where they cross the pass, however, is vague. In order to satisfy the outcrop pattern, and not to join faults with opposite displacement the relations shown on the map seem necessary. The fault starting at the klippe (north end of section E–E′, Pl. 2) curves westward and on its east side drops the Ogden thrust below the bottom of the canyon. This much seems definite, but the next fault on the east probably does not connect with either of the two faults across the pass to the south, because they have an opposite displacement. The fault on the north side has a considerable downthrow on the east, and therefore it probably does not end abruptly at the pass. If it curves westward and runs down the canyon to the first fault mentioned it satisfies the need for an east-west fault near the mouth of the canyon where limestone is opposite quartzite. A transverse fault is thus formed which has the appearance of a tear fault. Very little horizontal movement has occurred along it, as attested by the continuity of the structures across the pass, a short distance to the east. Also, the high-angle faults cut the latest thrust sheet, the Willard, and came into existence considerably later than the Taylor and Ogden thrusts. A small fault offsets the Tintic quartzite on the north side of Cottonwood Creek, and two small faults, close together, may be seen in the Weber quartzite in Weber Canyon above Morgan. A number of small faults are shown on Wegemann's map of the Coalville anticline. Their strike is normal to that of the beds, and their throw is just a few feet. Tear faults may be present beneath the Tertiary deposits in the Ogden Valley and Weber Valley areas, but none are suggested by the partially exposed structures.

SUPERPOSITION OF STRUCTURES

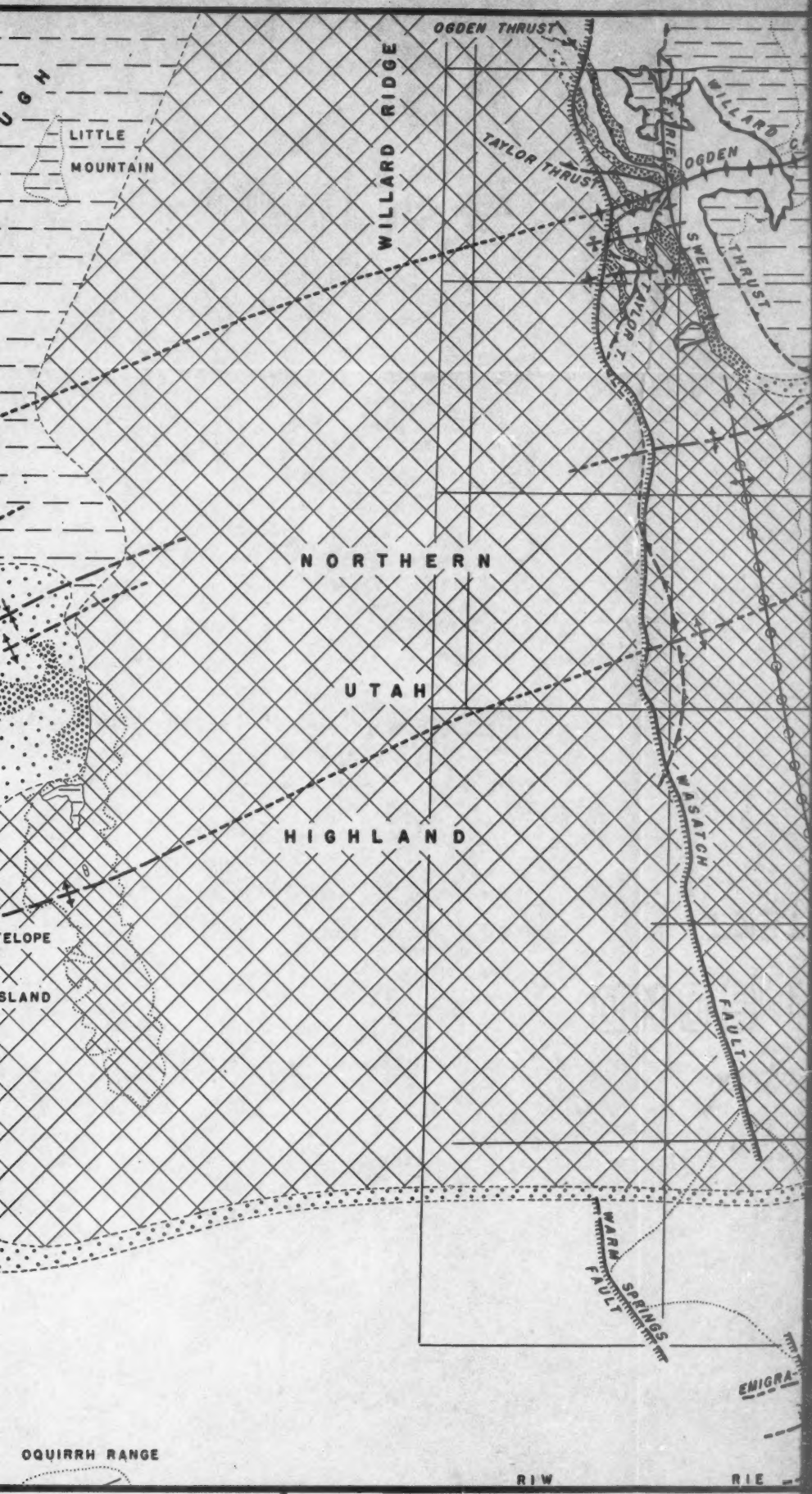
The Henefer anticline and the Stevenson Canyon syncline, both in Henefer and Wasatch sediments, follow along the projection of the folds of the central Wasatch. Since the Henefer formation rests unconformably on the underlying folded Mesozoic beds of the central Wasatch, and is itself folded, a second stage of deformation occurred in the already folded underlying beds, and the folds in the overlying beds approximately parallel those below. Of particular interest here is the superposition of the Henefer anticline over the Parley's Canyon syncline (Pl. 3). Perhaps this condition exists only in the immediate region of overlap. The unconformities of another fold, the Coalville anticline (Pl. 1, secs. D′–D″, E–E′, F–F′) show (1) a pronounced stage of folding in post-Frontier and pre-Henefer time; (2) that the unconformably overlying Henefer formation was gently folded in post-Henefer and pre-Wasatch time, but with an anticlinal axis slightly discordant in trend to the underlying Frontier axis and offset westward about 2 miles; (3) that the Pulpit conglomerate was upturned on the northwest flank of the anticline and, perhaps, elsewhere around it; and (4) that the unconformably overlying Saw Mill conglomerate was gently arched in the fourth diastrophic stage around the anticline. The axis of the gentle arch in the Saw Mill conglomerate over the underlying structure is impossible to fix exactly because the fold is so broad and so much of the conglomerate is eroded away.

The Coalville structure, the Stevenson Canyon syncline, and the Henefer anticline, together with the folds in the Twin Creek formation in Toone Canyon, are undoubtedly continuations of compressional structures from southeastern Idaho and southwestern Wyoming which veer westward as they approach the north flank of the Uinta uplift. The Coalville anticline seems to terminate abruptly against the great monocline of the north flank of the Uintas (Pl. 3), but a small area between has not yet been mapped in detail so the exact relation of the two is unknown. The Henefer anticline, however, seems to be continuous with the east-west folds of the central Wasatch which, in turn, are part of the north flank structure of the Uintas. The relatively simple though large monocline of the north flank of the Uinta uplift is complicated toward the west end by subparallel folds and faults where the structure crosses the Weber River (Pl. 3, T. 1 S., R. 5 E.), by the Park City depression, and by the folds of the central Wasatch. This is in part due to the proximity of the northern Utah highland. In all the above structures the elements of later origin followed approximately the earlier ones, but, commonly, fold axes were offset and in one place antithetical.

In the Ogden Canyon area the opposite is true—two groups of compressional structures cross each other at right angles. These are the cross folds and the thrust sheets. The cross folds fall in line with the folds on Promontory Point, Fremont Island, and the north end of Antelope Island (Pl. 3) which lie to the west, and these in turn line up with the north end of the anticline on Stansbury Island. Although connections cannot be established between individual anticlines and synclines because of the wide, separating cover of alluvium, the alignment suggests a continuous zone of structures. These folds, like those of the central Wasatch, appear to be continuations of the westward-veering structures of the Idaho-Wyoming arc. Most

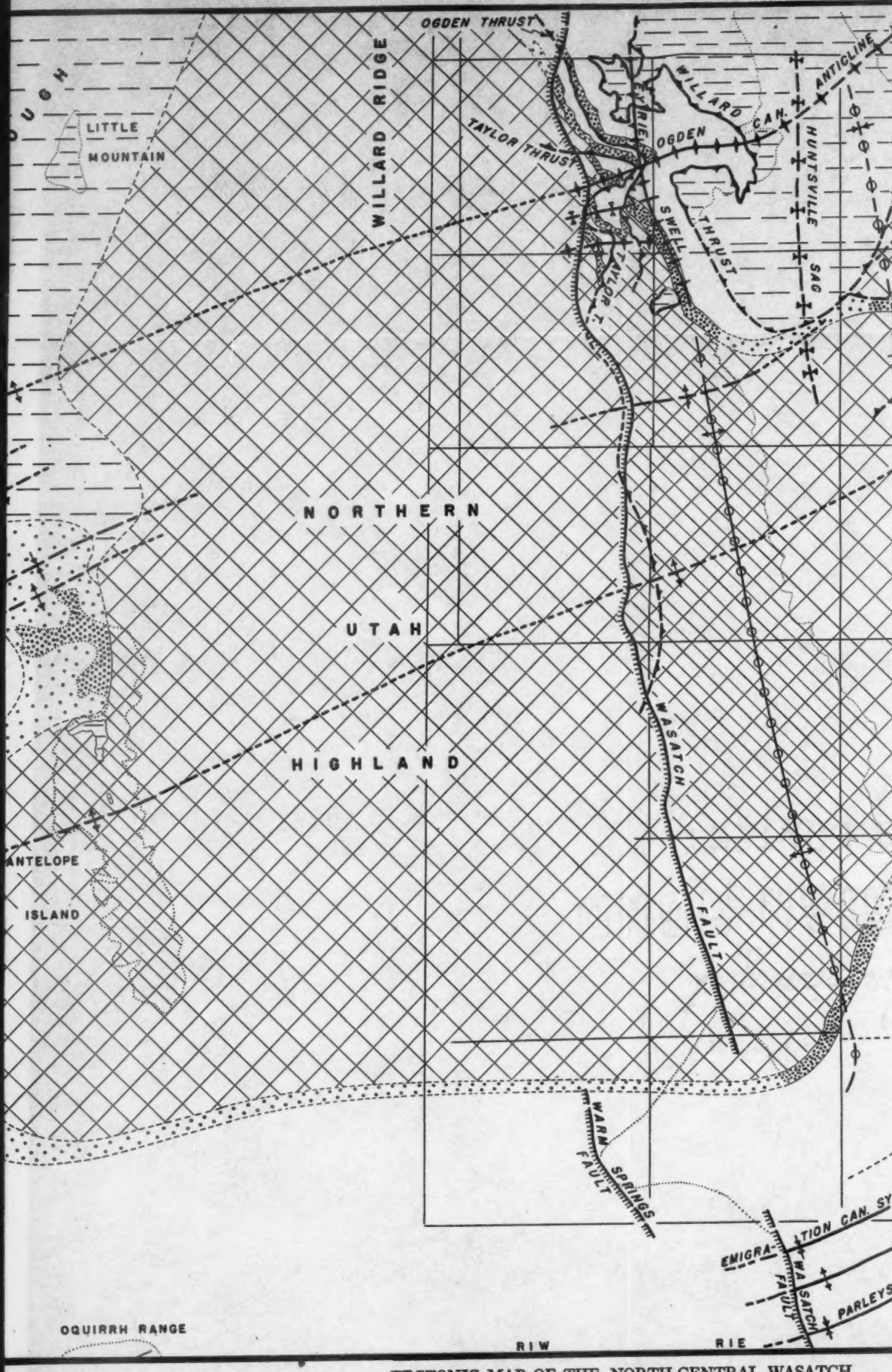


The part covered by the



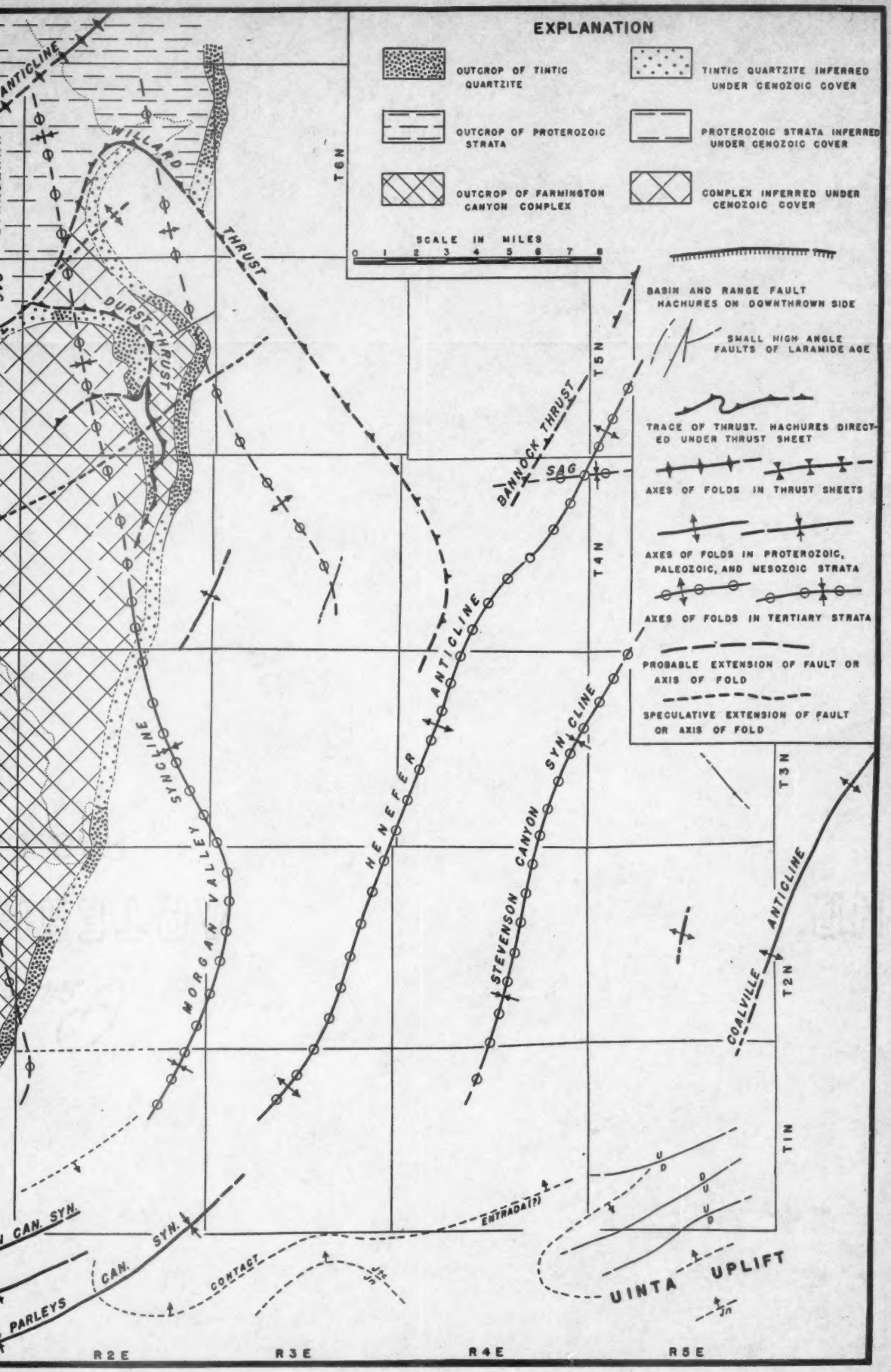
TECTONIC MAP OF THE NORTH-CENTRAL W

d by the township grid is the area of the geologic map, Plate 1. The area south of the grid is the c



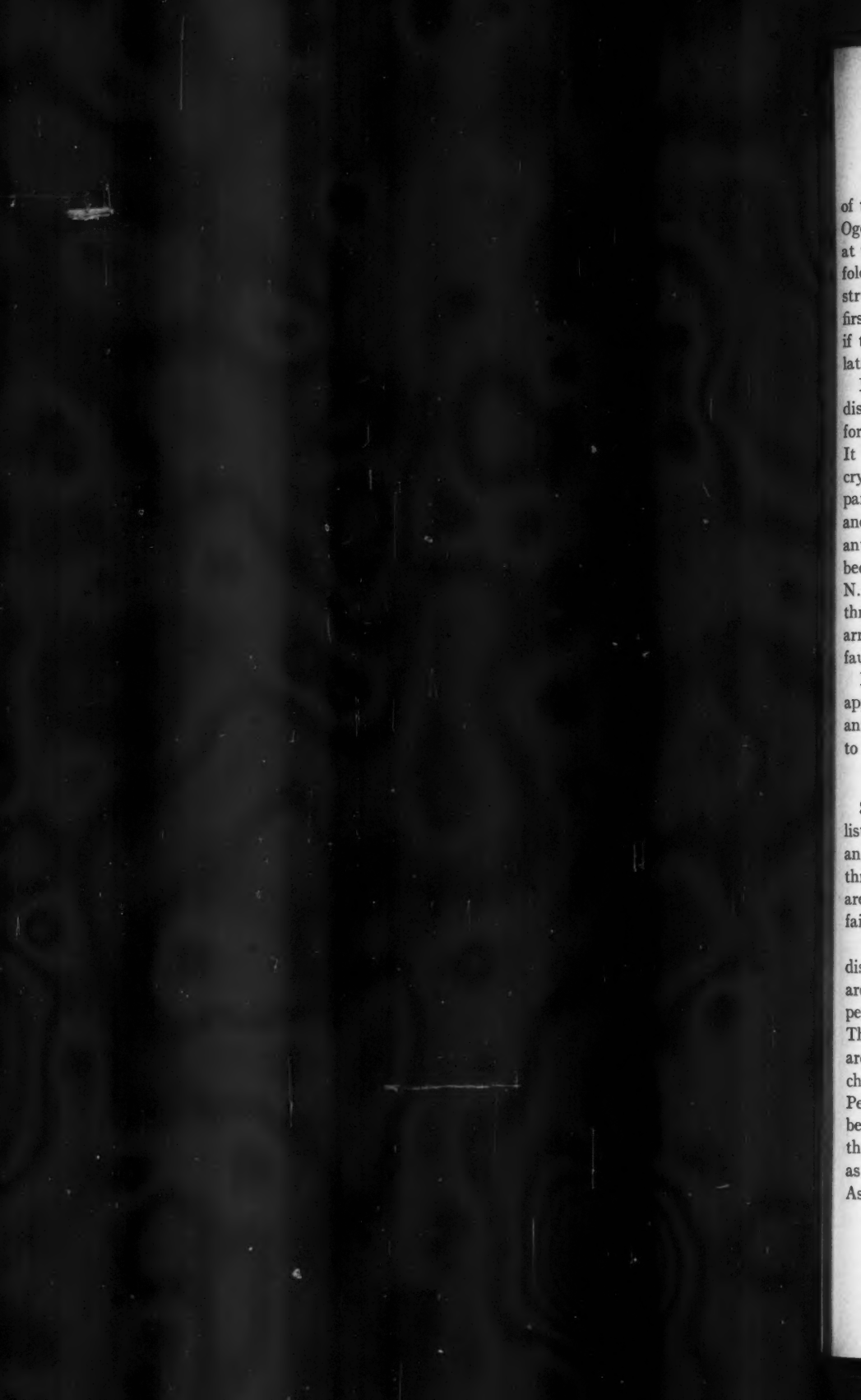
TECTONIC MAP OF THE NORTH-CENTRAL WASATCH

overed by the township grid is the area of the geologic map, Plate 1. The area south of the grid is the central Wasatch,



CH

Wasatch, and the area to the west is the Great Salt Lake Valley.



of the cross folding occurred before the Willard thrusting and after the Taylor and Ogden thrusting. The sequence of diastrophic events was: (1) upturning of beds at the highland margin and cross folding; (2) two stages of thrusting; (3) main cross folding; (4) thrusting; and (5) gentle cross folding. This succession of compressional structures indicates two sets of forces directed at right angles to each other, in which first one set and then the other manifested itself. Such a condition seems plausible if the local setting is viewed in its relation to the regional. This is done under a later heading.

In the closing stages of the Laramide revolution, gentle north-south folding was discordantly superposed on the older structures. The Morgan Valley syncline was formed in post-lower Eocene time and accentuated in post-lower Oligocene time. It is about continuous with the Huntsville sag in the Willard thrust sheet. The crystalline complex of the main frontal Wasatch became the core of a gentle anticline parallel with the Morgan Valley syncline (*see* section on Morgan Valley syncline), and the Durst Mountain "island" became the core approximately of another Tertiary anticline on the east. These gentle folds do not appear to be related to the thrusts because the folds trend only slightly west of north, whereas the thrusts trend about N. 40° W. The Tertiary folds depart widely in trend from the cross folds of the thrusts and the east-west folds in the central Wasatch. This new north-south arrangement of gentle folds marks the direction that the much later Basin and Range faults took.

Between Wolf Den Canyon at the mouth of Lost Creek and Toone Canyon, approximately along the site of Cedar Canyon, is a sag in the axis of the Henefer anticline. It may be related in trend to the cross folds of the thrust area and in time to the gentle cross folding that followed the Willard thrusting.

SEQUENCE OF DIASTROPHIC STAGES

Seven stages of compressional deformation in the Ogden Canyon area have been listed. Another sequence of diastrophic events can be recognized to the southeast, and it remains to enumerate them and relate them to the stages of the Ogden Canyon thrust structure. Since the diastrophic stages in the southeastern part of the area are revealed in a number of unconformities, the ages of the related sequences can be fairly well determined.

The coarse conglomerates of the Kelvin (?) formation mark the first major crustal disturbance of the general Sierran-Laramide revolution in the north-central Wasatch area. The pebbles of the lowest conglomerate are all chert, but they decrease in percentage in successively higher conglomerates until quartzite pebbles predominate. This occurrence is similar to others found by Stokes (1941, p. 48, 134) in adjacent areas. He has identified Mississippian, Pennsylvanian, and Permian fossils in the chert pebbles and recognizes a lithology in the quartzite pebbles that resembles the Pennsylvanian quartzites of the Weber quartzite and the Oquirrh formation. He believes (personal communication) that the Permian cherts were widely exposed in the earliest uplift and supplied most of the pebbles of the lowest conglomerates, and as erosion cut deeper the Pennsylvanian quartzites became the dominant source. As the coarsest conglomerates are found at their westernmost exposures in Emigra-

tion Canyon (in the central Wasatch—not on map) the source, and therefore the area of uplift, lay to the west. The boundary between the area of uplift and the area of accumulation was very close, evidently, to the present Wasatch front.

Upturning of beds at the eastern margin of a western highland and their cross folding were described as the first stage of deformation in the Ogden Canyon thrust area. It is possible that this is the same stage as the one that resulted in the orogenic deposit, the Kelvin conglomerate.

The early uplift of the western highland was either in the closing stages of the Jurassic period or in lower Cretaceous time, but the conglomerates have not yet been accurately dated. The disturbance is probably closely related to the Sierran revolution which was much more intense and consisted of at least two episodes several hundred miles to the west.

Mansfield writes of an uplift in the region west of the principal deposits of the Gannett group in Idaho, an uplift that seems closely related to the Kelvin (?) uplift of the north-central and central Wasatch. In the uplifted region he states that the rocks exposed were principally Paleozoic.

The next major diastrophic event that can be dated is the nearly east-west folding in the central Wasatch Mountains between the northern Utah highland and the Cottonwood uplift. Here, the rocks exposed in the folds are Paleozoic and Mesozoic up to and including the Frontier beds, and the earliest strata in angular unconformity above them are those of the Henefer formation. The structure of these folds has been traced by R. E. Marsell (personal communication) continuously into the north flank of the Uinta uplift. (*See* Plate 3.) Thus a post-Frontier and pre-Henefer disturbance marks the earliest important uplift at the northwest end of the Uintas. This corresponds to the stage of major folding on the south flank of the Uintas recently described by Walton (1944, p. 126). The magnitude of the disturbance can be judged by the fact that the Henefer overlies the truncated edges of at least 25,000 feet of strata (Woodside to and including Frontier), and it may actually have been deposited over the Paleozoic rocks as well.

The Henefer formation is not exposed in contact with the Willard thrust sheet, but near Croyden, toward which the buried trace of the Willard thrust is projected, the formation is so sharply and asymmetrically folded that it probably predates the late thrusting. If the Willard thrusting followed a long interim of erosion, which separated it from the earlier thrust sheets and cross folds, as seems proved, the Taylor thrusting, the Ogden thrusting, and the main part of the cross folding should correspond in age to the central Wasatch folding and the first major Uinta uplift. Since the cross folds of the Ogden Canyon area parallel the central Wasatch folds it seems logical, although not necessary, that they relate specifically to the central Wasatch folding and that the Taylor and Ogden thrusts came into existence somewhat earlier, but still in post-Frontier time.

The interim of erosion that followed the early thrusting and cross folding in the Ogden Canyon area would correspond to the stage of erosion in the central Wasatch that followed folding and, in addition, to the time during which the Henefer formation was deposited. It is believed that the two early thrust sheets became part of the margin of the western highland and were subjected to erosion almost continuously until covered by the Willard thrust sheet.

The time during which the Upper Cretaceous beds were being deposited (Frontier and Aspen (?)) needs to be accounted for. The subsidence of a trough deep enough to receive at least 10,000 feet of sediments is a diastrophic event that must not be neglected. The highland to the west probably continued to rise after the Kelvin (?) uplift and furnished most of the upper Cretaceous sediments. The highland was either low or quite distant during most of Aspen (?) and Frontier time because the sediments of these formations are generally fine. The coarse conglomerate above sandstone No. 3 evidently indicates an exceptional condition. Spieker and Reeside (1926, p. 429–438) have described shore lines of a highland southwest of the Uintas during Upper Cretaceous time and suggest that the highland extended northward into the area of this report.

Veatch (1907, p. 64) observed that the Bear River formation in western Wyoming thins and dies out southward toward the Uintas. This is an indication, he believed, that the Uintas had begun to rise at that time.

The bentonite and porcellanitelike beds in the Aspen indicate an episode of volcanism, although the center was probably some distance away. This was not the first volcanism, however, because a green ash bed is found in the lower part of the Twin Creek formation, and bentonitic shales are recognized in the Kelvin (?).

A 200-foot unit of tuff also occurs in the Henefer formation. Thus far, four volcanic stages in the general Sierran-Laramide revolution in the north-central Wasatch have been noted.

As already inferred the Willard thrusting occurred next. The Henefer anticline and Stevenson Canyon syncline in the Henefer formation were formed at this time. The Coalville anticline, in which the Frontier beds were folded in pre-Henefer time, was compressed, and the overlying Henefer beds were arched gently, but with a slightly offset axis. Since the central Wasatch folds including the Coalville anticline had already come into existence, their parallelism to the Henefer anticline and Stevenson Canyon syncline suggests that the older folds control the younger and that the older folds continue under the Henefer formation to the northeast for a number of miles.

Extensive erosion followed the Willard thrusting. The Henefer formation was eroded from the Coalville anticline, and the Frontier beds were re-exhumed. By tracing this erosion surface northwestward under the Almy conglomerates to the crest of the Henefer anticline (*see* section D′–D″ and F–F′, Pl. 1) it is seen there that at least 8000 feet of Henefer beds was truncated. The Willard thrust sheet itself was deeply and widely dissected. The great Huntsville-Morgan Valley, site of the present Morgan Valley syncline, had been cut to within 1000 or 2000 feet of the bottom of the Willard thrust sheet in Ogden Valley, by the time the Almy conglomerates had begun to accumulate. The adjacent mountains were not very high, however, and the valley was about as wide but probably not so deep as that existing today. The area to the northeast, now a dissected plateau underlain by Wasatch beds, was perhaps a plains country and the future site of extensive alluviation.

The deeply eroded Willard thrust sheet became buried in part by the strata of the Wasatch group. The Pulpit conglomerate, lowest of the group, covered an extensive area east of Durst Mountain and north of the Coalville anticline. The pebbles and boulders of the Pulpit in the local area studied were probably all derived from the freshly uplifted western highland, but conglomerate beds of similar age probably

flank considerable areas at the northwest end of the Uintas and may have come in large part from that range. This would indicate a second disturbance in the Uintas immediately preceding or during Pulpit time. Walton (1944, p. 127) does not recognize a second uplift in the Uintas until pre-Uinta time, but the absence of the Wasatch and Green River beds in his area represent a fairly long hiatus.

The Pulpit conglomerate was then folded sharply on the northwest side of the Coalville anticline, and the upturned beds were truncated. More boulders were swept into the local area, still chiefly from the west, and the Saw Mill conglomerate formed over the eroded edges of the Pulpit conglomerate and the Henefer and Frontier formations. The accumulation of conglomerate was interrupted by a fourth episode of volcanism, and 800 feet of ash interlayered and mixed with sand and clay was deposited. The beds accumulated in a lake of uncertain extent. Conglomerate again began to accumulate, now principally in the Morgan Valley area. A wide valley existed at this time and was filled to overflowing. The area of Durst Mountain was covered, and the low crest of the frontal Wasatch from Bountiful Peak to the Salt Lake salient was also blanketed. Because of the extensive cover of later sediments the amount of conglomerate that accumulated west of the present frontal Wasatch in the valleys beyond is not known. However, a large body of conglomerate occurs in the Salt Lake salient, west of the range proper. This is probably the Knight (?) formation and has considerable sand and shale, and boulders of very local derivation.

Next was a stage of gentle folding of the Wasatch beds. The Morgan Canyon syncline had its inception, and the flanks of the Henefer anticline, the Stevenson Canyon syncline, and the Coalville anticline were slightly steepened. The Almy conglomerate was flexed locally just north of Croyden in the saddle of the Henefer anticline so as to dip 50°.

A second cycle of valley erosion began even before the post-Wasatch folding. The erosion continued during and after the folding and ended in the establishment of a deep valley in the Henefer anticline. Here the Almy conglomerate was extensively stripped, and the wide valley bottomed in the re-exhumed Henefer formation about 2000 feet stratigraphically below the Wasatch.

The main part of the Park City and Kamas Valley volcanic eruptions occurred at this time, and much of the fragmentally ejected material was washed northward into large valleys to form the Norwood tuff. Although drainage was across the frontal Wasatch at this time, the distribution of bedded volcanic sediments west of the Wasatch is so extensive that additional sources of volcanic material might be inferred, and it seems probable that they represent several post-Wasatch episodes of volcanism.

The Norwood tuff was gently folded in the Morgan Valley syncline as the flanks of the older syncline were steepened. Dips along the west side of the valley in the Norwood tuff correspond approximately to dips in the underlying conglomerate, and the same is true on part of the east flank in Norwood Canyon. The gentle folding also resulted in the elevation of Durst Mountain and the frontal Wasatch as previously described. The Willard thrust sheet was also gently folded at this time in a north-south direction (Pl. 1, sec. A–A′; Pls. 3, 5), and the broad folds in it have been called the Eyrie swell and the Huntsville sag.

Bradley (1936, p. 167) has pointed out that the Eocene deposits near the fault and flexure of the north flank of the Uinta uplift have been deformed. On the south flank the major unconformity is between the lower Oligocene Duchesne River and the Eocene Uinta (Walton, 1944, p. 127), and the uplift that this indicates corresponds to the pre-Norwood tuff disturbance in the north-central Wasatch. Another gentle uplift of the Uintas occurred, according to Walton, in post-Duchesne River time. This corresponds to the folding of the Norwood tuff in the north-central Wasatch.

Following the folding of the tuff two extensive erosional surfaces developed. The cycle of the second erosion surface was interrupted by Basin and Range faulting.

In summary, a succession of eight diastrophic stages representing the combined Sierran-Laramide revolution can be recognized in the north-central Wasatch. Five episodes of volcanism and seven erosional cycles are also noted during this time. The events took place between Upper Jurassic and Upper Oligocene time and are summarized in Figure 2.

RELATION OF LOCAL STRUCTURES TO REGIONAL

Time relations.—The local structures may be related to the regional by correlating the deformational stages of local areas with each other and by comparing the local structural pattern with that of the regional. The two are interwoven in places because in the succession of deformational stages one structure has been superposed discordantly on another. The correlations are limited to the geosynclinal belt along the northwestern margin of the middle Rocky Mountain province, because (1) The correlations deal with a single tectonic unit, and (2) sufficient geologic data are now available from this region to make the attempt significant. Figure 2 has been prepared to show the time relation of diastrophic events in the geosyncline. Because of their attention to diastrophic detail the reports of G. M. Mansfield (1927) and J. D. Love (1939) have been used as sources of information about the region north of the area studied by the writer. It would be highly desirable to include E. M. Spieker's work in central Utah, but only an abstract is in print, and he has done considerable work since the abstract was published. The epoch and "provincial" age classification was taken from the recent report of Wood *et al.* (1941). Where disturbances are dated in relation to formations, the formations and thus the disturbances were related to the "provincial" ages as set forth in the above report. The per cent marks opposite Jurassic, Cretaceous, Tertiary, and Quaternary indicate the amount of time assigned to these periods by Goodman and Evans (1941, Fig. 10, p. 528). By means of these the attempt is made to represent time and thus the proper spacing of geologic events. The Lancian epoch, recently proposed by Dorf (1942, Fig. 3, p. 110), between the Montana and Paleocene is represented by the upper Montana in the diagram. If inserted as a separate epoch, either or both the Montana or the Tertiary would have to be shortened. By its nature a time and space correlation diagram demands a greater definition of opinion than may be justified, but question marks indicate the approximate age or nature of the assignments. The curves undoubtedly represent only a partial picture of the deformational stages of the geosyncline, but nevertheless a few general conclusions can be drawn from the chart.

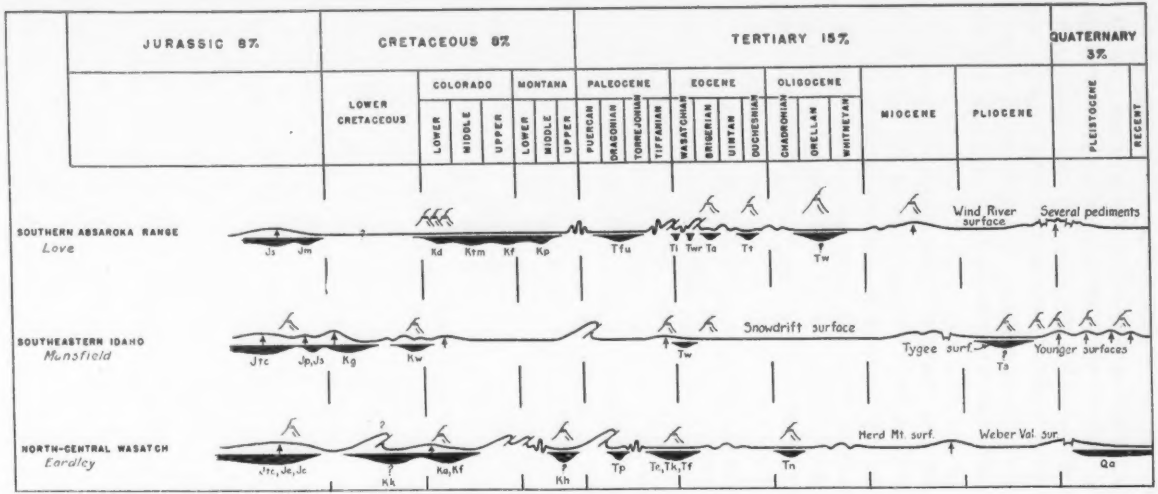


FIGURE 2.—*Correlation diagram of diastrophic episodes*

Diagrams correlate events in three parts of middle Rocky Mountain geosyncline—north-central Utah, southeastern Idaho, and west-central Wyoming. Heavy lines represent manner and time of crustal disturbance, and black lenses under lines, the adjacent or associated basins in which sediments accumulated. Broad arch with arrow under it indicates epeirogenic uplift, open folds in lines indicate gentle or broad folding, isoclinal folds in lines indicate fairly intense folding or creation of folds whose flanks dip 30° or more, thrust diagrams indicate thrusting, and grabens indicate high-angle faulting. Volcanic cones indicate times when volcanoes were active either in the area or near by. Symbols for formations of southern Absaroka Range are as follows: Js, Sundance; Jm, Morrison; Kd, Dakota group; Ktm, Thermopolis and Mowry; Kf Frontier; Kp, post-Frontier; Tfu, Fort Union; Ti, Indian Meadows; Twr, Wind River; Ta, Aycross; Tt, Tepee Trail; Tw, Wiggins. For southeastern Idaho: Jtc, Twin Creek; Jp, Preuss; Js, Stump; Kg, Gannett; Kw, Wayan; Tw, Wasatch; Ts, Salt Lake. For the north-central Wasatch: Jtc, Twin Creek; Je and Jc, Entrada and Curtis; Kk, Kelvin; Th, Henefer; Tp, Pulpit; Ts, Sawmill; Tf, Fowkes, Tk, Knight; Tn, Norwood. Dating of events in Absaroka area (including Bighorn Basin) in Paleocene time was facilitated by reference to Dorf (1940, Table 5, p. 230), Love, (1939), and Mansfield (1927).

In latest Jurassic or Early Cretaceous time conspicuous unrest in the highland west of the Mesozoic geosyncline was manifest. The highland rose sharply along a meridian marked now by the western front of the Wasatch Mountains, and from it the basal conglomerates and associated beds of the Kelvin (?) and its correlatives were derived. The conglomerates are so coarse and voluminous that deformation by thrusting has been visualized by Spieker (personal communication). When stratigraphic correlations are better established for this interval of time several impulses of deformation may be recognized. The eastern margin of the highland was cross folded in the north-central Wasatch area. Distant volcanism both preceded and followed this early period of unrest.

Most of Lower Cretaceous time, it seems, and the Colorado epoch of the Upper Cretaceous elapsed with erosion of low-lying uplands and subsidence and deposition. The early part of Colorado time in the geosyncline was especially one of subsidence.

The second main group of compressional impulses began in late Colorado or early Montana time and reached a climax generally in late Montana and early Paleocene time. The Taylor and Ogden thrusting of the north-central Wasatch seems to have started off the succession, then folding of the north flank of the Uintas with associated east-west folding in the central Wasatch and cross folding of the Taylor and Ogden thrust sheets continued. Episodes of folding, and folding and thrusting, south of the Uintas also occurred at this time. Following upon this came the great Willard and Bannock thrusting of northern Utah and southeastern Idaho. Distant volcanoes were active again in late Paleocene, with folding recorded in central Utah.

The central sector of the geosyncline appears to have undergone its principal deformation by this time and subsequently suffered, in succession, gentle folding and uplift, volcanism, and more gentle folding, the last of which terminated probably in upper Oligocene time. To the north, however, in the Owl Creek and Absaroka Mountains, the main deformation just began. In early and early middle Eocene, Love records five impulses of folding and thrusting. Late thrusting also occurred farther north along the Absaroka Mountains in the Shoshone Reservoir area. Pierce (1941, p. 2021–2045) records two stages of thrusting in lower Eocene. South of the central sector of the orogenic belt Spieker has found that intensive deformation in the form of monoclinal folding, normal faulting, local folding, and thrusting occurred during the Eocene. The third principal group of compressional impulses evidently occurred, therefore, in early, middle, and late Eocene time. Active volcanoes were widespread during the Oligocene and Miocene and emitted great volumes of extrusive rock both in the northern and southern regions.

Miocene and Pliocene time was mainly one of erosion and epeirogenic uplift. Although dating is not yet very reliable, faulting of the Basin and Range disturbance in the northern Wasatch probably began toward the end of the Pliocene. Most of it was completed by middle Pleistocene time although more recent movements along some of the faults have been noted. In place of Basin and Range faulting along the southern margin of the Absaroka Range, regional uplift was dominant with local folding and faulting. Spieker (personal communication) has recognized Pliocene or Pleistocene thrusting in southern Utah and possibly in central Utah. Near the Hurricane fault the thrusting post-dates the last movement on the fault.

Spacial relations.—In a previous report the writer (Eardley, 1939, p. 1294–1298) indicated the major elements of the regional pattern and pointed out that cross folding of the thrust sheets probably had occurred. Now it seems evident that some of the folds of the Idaho-Wyoming arc continue across the thrust sheets and also

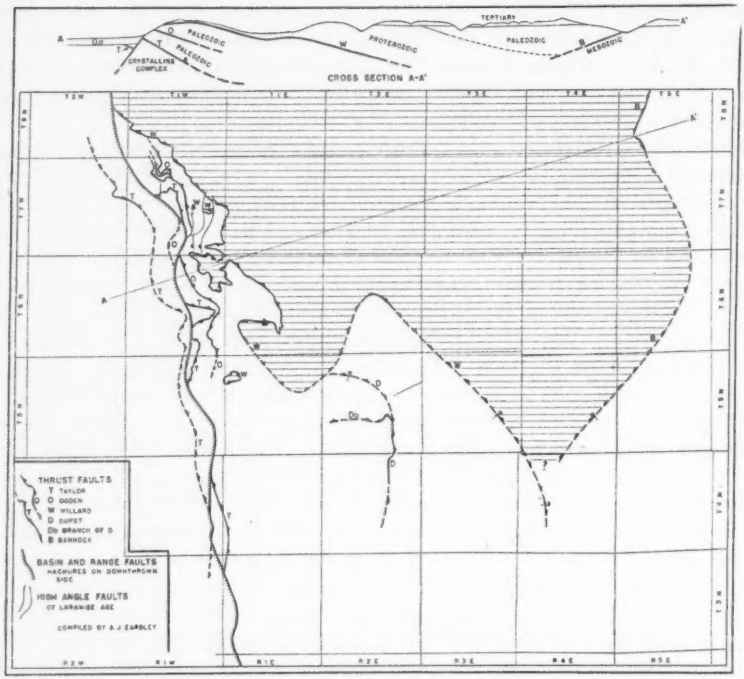


FIGURE 3.—*Fault map of north-central Utah*

across the northern Utah highland. In Plate 3 the first interpretation of the northern Utah highland is amplified by adding the data collected since.

The complex north flank monocline of the Uintas projects westward into the folds of the central Wasatch. A continuous orogenic belt is formed by the two structures, but as the folds of the central Wasatch are traced to the northeast they depart entirely from the monocline and become part of the Idaho-Wyoming arc. The Coalville anticline seems exceptional and terminates abruptly at the monocline without cross folding of the monocline or conspicuous turning to the west.

The relation of the thrusts of the north-central Wasatch to the Bannock is shown on Figure 3. Richardson (1941, Pl. 1) has traced the Bannock overthrust of Idaho southward into Utah across the Randolph quadrangle to the headwaters of Lost Creek, just 12 miles north of the northern border of this report. The blanket of Tertiary conglomerate is not complete, however, and when the outcrops in this inter-

vening area are studied and mapped more light will be shed on the relation of the Bannock thrust to the structures of the north-central Wasatch. It appears at present that the thrust is deflected westward and enters the northern part of the area of this report, in harmony with the trend of the Henefer anticline. Since the fault is believed to be early Paleocene (Mansfield, 1927, p. 199) the Henefer formation was probably deposited before the Bannock thrusting occurred, and hence the Henefer sediments would not be unconformable on, and masking the trace of, the thrust. The thrust, therefore, would have to lie west or east of the visible core of the Henefer anticline. If to the west, as seems probable, it will about meet the Willard thrust in the vicinity of Croyden. The two great thrusts dip toward each other, and the strata between them may be complicated by imbricate thrusts and tears, but unfortunately the area is mostly covered by gently folded conglomerates, and the underlying structures are not discernible.

The cross section at the top of Figure 3 does not show the manner in which the great thrusts connect underground because the relations are not known. Two hypotheses will be presented, and in order to make clear the three-dimensional meaning of Figure 3, before the problem is further discussed, the reader is referred to a perspective drawing, Plate 5. This shows the thrusts only down to a sea-level plane.

The hypothesis that the writer favored in previous papers is that the Willard is a westward overthrust and the Bannock an eastward. It is illustrated as the orogen in Figure 4. According to the other hypothesis, the Willard connects with the Bannock and with it forms a great eastward overthrust that is rooted many miles west of the Wasatch (Fig. 4).

The support for the hypothesis of the orogen is as follows:

(1) The Willard thrust surface dips northeastward, as do the underlying Taylor and Ogden thrusts, although not so steeply. The lower sheets stand opposed to eastward movement of the overriding Willard.

(2) The Taylor and Ogden thrusts have dragged beds westward beneath them. Drag under the Willard sheet is displayed in Willard Canyon (Eardley and Hatch, 1940, p. 810, Fig. 3) and, although not conclusive, suggests westward movement.

(3) It would be exceptional if two earlier thrusts moved westward, and then a later one moved eastward over them.

(4) The stratigraphic displacement and apparent horizontal displacement of the Willard is much greater than the Bannock in the line of the cross section, Figure 3. If the Willard and Bannock are one sheet, the front of the Bannock would have as much horizontal displacement as any other part and probably a greater stratigraphic displacement. Normally the greatest erosion should take place where the oldest rocks are exposed. Nowhere along the Bannock are rocks older than Paleozoic exposed. Also, the diagram illustrating the hypothesis of eastward thrusting is impossible because the Eyrie Peak part of the thrust sheet moved much farther than the Bannock part. If the hypothesis is true, the diagram will have to be changed either by removal of the Mesozoic on the thrust front or by constructing another thrust slice in imbricate arrangement with the Bannock, probably above it,

and assuming that most of the movement took place on the imbricate fault. Such a fault could exist beneath the extensive Tertiary cover, but so far no tangible surface indications of it are known. The conditions near the Bannock thrust, necessary to make the interpretation of eastward thrusting possible, are shown in thin dashed lines on the lower cross section of Figure 4.

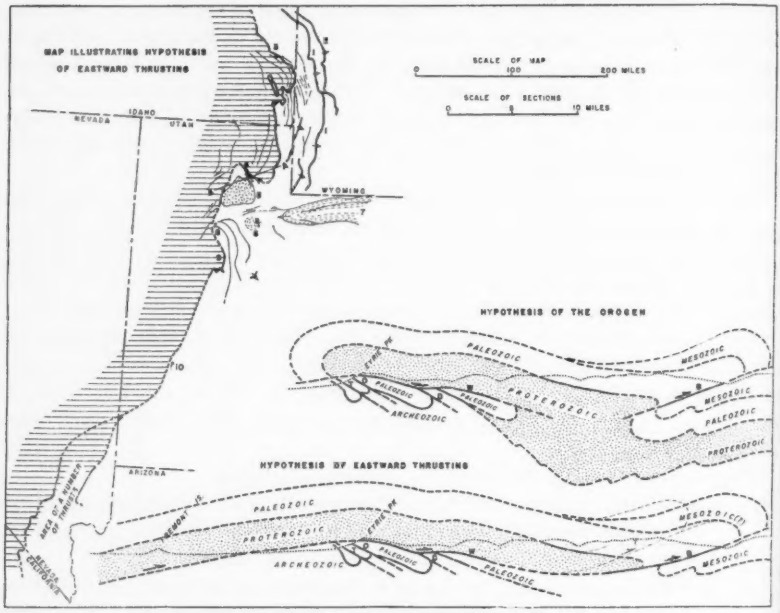


FIGURE 4.—*Hypotheses of thrust structure*

Hypothesis of the orogen envisions westward thrusting of the Willard and eastward of the Bannock. Hypothesis of eastward thrusting of both requires roots of a great sheet in Fremont Island and the area west of Antelope Island. The accompanying map shows a possible position of such a thrust trace. Lower cross section is along line A-A′. Numbers on map: (1) Absaroka thrust, (2) Darby thrust, (3) Bannock thrust, (4) Willard thrust, (5) northern Utah highland, (6) Cottonwood uplift, (7) Uinta uplift, (8) South Mountain, (9) Sheeprock thrust, (10) San Francisco Mountains. Letters on cross sections: (T) Taylor thrust, (O) Ogden thrust, (W) Willard thrust, (B) Bannock thrust. Compare with Eardley (1939, Pl 1, p1 1282) and Nolan (1943, Fig. 12, p. 175).

(5) In the region of supposed roots, the contact between the Proterozoic and Archeozoic at the one place exposed, Antelope Island, is sedimentary, not thrust, and the Proterozoic island sequences have been correlated on the basis of sedimentary overlap. The structure of the cross section showing the correlation (Eardley and Hatch, 1940, p. 800, Fig. 2) is, however, largely hypothetical.

(6) No great thrust has been described west of the Wasatch with which the Willard might connect. A structural pattern was proposed by the writer (Eardley, 1939, Pl. 1, p. 1282), which, although admittedly fragile in parts, showed no indication of a great thrust such as diagrammed in Figure 4.

(7) The greatest obstacle to the acceptance of the hypothesis of eastward thrusting has been the size of the postulated structure. The writer seemed inhibited from giving the hypothesis a fair consideration because the distance between salients and recesses of the imagined thrust front would be at least 45 miles (map, Fig. 4), and horizontal displacement could hardly have been much less and may have been more. This seemed until now a potent reason for favoring the hypothesis of the orogen.

The observations and deductions that favor the hypothesis of eastward thrusting are as follows:

(1) A thrust as great as the Willard cannot die out within 20 or 25 miles. The hypothesis of the orogen embodies the concept of rather abrupt termination. Inspection of Plate 5 shows that the Willard sheet has great width over the north-central Wasatch Mountains, and then before reaching Croyden it is very narrow. If the general strike of the Willard is projected northwestward only 25 miles, a large area of Mississippian and Pennsylvanian rocks lies athwart its course, and there seems to be no place for the thrust in this direction.

(2) It is difficult to relate the central Wasatch and Uinta folds to the postulated orogen. The Willard and Bannock are underthrusts according to the hypothesis, and the soles were the active elements which moved in and under. If so, the rocks under the Willard thrust should be very much deformed, and the belt of deformation should extend southward. The Henefer beds should be strongly folded and thrusted, but only the Henefer anticline is known between Croyden and the central Wasatch.

(3) If the hypothesis of eastward thrusting has any merit an area of roots should exist to the west. The Proterozoic sequence on Fremont Island in the center of the postulated Fremont trough (Pl. 3) has the varved phyllites, the tillites, the marbles, and many of the quartzites that make up the Willard thrust sheet. Were it not for the fact that the Cottonwood uplift has similar lithologies (map, Fig. 4) the identity of the rocks in the Willard thrust sheet with those of the Fremont trough would be convincing evidence of large-scale eastward thrusting. It seems impossible to pass the thrust fault east of the Cottonwood uplift and include the Proterozoic rocks of the uplift in the great thrust sheet, because the geology will not allow such an interpretation. The similar lithology of the Cottonwood Proterozoic rocks need not be very disturbing, however. The Proterozoic rocks of the Uintas and those of the Cottonwood uplift appear to have formed in a separate trough running eastward from the Fremont (Eardley and Hatch, 1940, p. 830, Fig. 6). Although the two sequences have characteristics in common, they are postulated to have formed some distance apart, and therefore the western could have been thrust over part of the eastern.

(4) If such a great thrust sheet exists, not only should its roots be found to the west, but it should have a length of hundreds of miles, and its trace should extend for a long distance southward. Even though the geology of most of western Utah is known only in reconnaissance, it should be possible to

find a logical position for the trace without serious difficulties. By utilizing the areal geologic maps of the State an attempt proved encouraging. The line on the map of Figure 4 connects the Willard with the Sheeprock thrust in central Utah and the Sheeprock with a number of large thrusts in southern Nevada, so that fairly dissimilar rocks or discordant structures are separated from each other by it. The trace would pass under the alluvium between Fremont and Antelope islands so that the Proterozoic sequence of Fremont Island would be in thrust contact chiefly with the crystalline complex of Antelope Island. The complex was left veneered in places with a thin cap of upper Proterozoic and Cambrian strata. If the trace of such a thrust exists, the relation of Fremont Island to Antelope Island illustrated on Plate 3 and in a previous publication (Eardley and Hatch, 1940, Fig. 2, p. 800) would not hold. The trace would continue southwestward between Antelope and Stansbury islands, still under the Great Basin deposits, and thence southward along the east front of the Stansburg Range to and through the pass between South Mountain and the Stansbury Range. The trace between Stansbury Island and Antelope Island runs under alluvium that separates Carboniferous rocks from Archeozoic, but between the Stansbury Range and the Oquirrh Range Paleozoic strata are on each side. A discordance and conspicuous break between the two ranges were not recognized in a previous study, but now the relation of the two ranges must be reexamined. The Laramide folds pass northwestward and westward through the Oquirrh Range and north-northeastward through the Stansbury. The east-west fold of South Mountain (Gilully, 1932, Pl. 1) is overturned northward, and previous to the preparation of the regional structural map of the writer's earlier report (Eardley, 1939, Pl. 1) the attempt was made to trace the South Mountain overturned fold into the Stansbury Range. A perplexing assemblage of volcanic rocks and shattered limestone outcrops was found. The location is marked by a question mark on Figure 4. It was concluded that detailed mapping would be necessary to make out the structure. Along the crest of the range, however, an anticline with Cambrian strata in the core trends north-northeast. Although Paleozoic rocks occur on both sides of the hypothetical thrust trace, the evidence does not prohibit the existence of a thrust and, as far as known, can imply structural discontinuity, as in Figure 4, or continuous folds, as in Plate 1 of the earlier report.

From South Mountain the thrust would run under the alluvium of wide Rush Valley and join the great Sheeprock thrust in the Sheeprock and Tintic ranges. There the Proterozoic strata are thrust over deformed Paleozoic rocks in the same manner as in the north-central Wasatch. The similarity in the structure impells joining the two areas as parts of the same great overthrust. The route described for the trace of the thrust from one area to the other is the only one that the reconnaissance geologic maps seem to permit.

From the Sheeprock uplift southward the alluvium of the broad Sevier Desert, and the lava fields west and southwest of Filmore, conceal all bedrock in the area, and somewhere under these deposits the thrust would have to pass. The position is narrowed down to the strip of alluvium between the San Francisco Mountains on the west and the Beaver Lake Mountains on the east, and farther south between lower Paleozoic outcrops on the west and Triassic on the east. This area is in Beaver County, Utah, and the hypothetical line would run west of all Triassic deposits and separate them from rocks of early Paleozoic age.

From Beaver County the trace would veer westward under the alluvium and volcanics of the Escalante Desert into Nevada and connect with any or several of a number of thrusts in southernmost Nevada (Nolan, 1943, Fig. 12). Under Escalante Desert the thrust would separate the highly deformed Paleozoic rocks from the gently folded Mesozoic and Paleozoic rocks under the extensive volcanics of western Washington County, Utah.

(5) The trace of the thrust thus delineated shows a maximum distance between the salients and embayments in the thrust front, measured normal to the general strike, of about 45 miles. The window is about 32 miles from the front of the Bannock thrust sheet. The two thrusts, therefore, have somewhat similar proportions. Great stratigraphic displacement is very significant where found, but lack of it does not mean, necessarily, any less horizontal displacement.

(6) In working out the evolution of a great thrust structure the deposits on the foreland are commonly as meaningful as the structures of the hinterland. The concept of flysch and molasse type sediments illustrates the point. In the foregoing pages the sequence of deformational events is fairly well established, but the spacial relations of the orogenic deposits to the episodes of thrusting has probably proved confusing to the reader. The problem is complicated by the Uinta arch which rose in the Paleozoic and Mesozoic foreland and which may have been the source of some of the sediments. The picture would be less complicated, certainly, if all the mountain building episodes were in a western hinterland, and all the deposits were transported to eastern piedmonts and basins. The hypothesis of eastward thrusting lends itself to this version. Accordingly, the Kelvin, Henefer, and Wasatch formations would represent three great uplifts of the western hinterland, and the pebbles and boulders of these formations in the limited area studied confirm the concept. The Henefer formation may have been overridden in part by the Willard sheet. Whereas the Willard thrusting has been dated as post-Henefer, now it appears, according to the hypothesis of eastward movement, that as the thrusting started to the west the Henefer clastics began to accumulate on the east, and that the thrusting continued until the deposits were eventually overridden. The writer would be satisfied with this relation of sediments and crustal movements if it were not for the evidence of westward thrusting of the Taylor and Ogden sheets.

The two hypotheses will be left standing on their strong and weak legs, such as the writer has been able to build for them, and should serve to focus future structural and stratigraphic work in the region.

PHYSIOGRAPHY

LATE LARAMIDE SURFACES

A number of interims of erosion separated the deformational stages of the general Sierran and Laramide revolution in the north-central Wasatch, and what little is known about the early erosion surfaces up to the deposition of the Norwood tuff in lower Oligocene time has already been related. The surface upon which the Norwood tuff accumulated may be considered from two extreme points of view. Either wide, deep valleys existed, in the position of and much like those of today, in which the volcanic material was deposited and later folded, or the tuff was spread on a surface of low relief, folded, and then eroded from the anticlines. Since no remnants of tuff are found in the Weber Valley from Henefer to Wanship, this particular depression did not exist then as now. The main drainage from the Park City and Kamas volcanic field may have followed down a depression where East Canyon above the reservoir is now located, and then northward to the position of Little East Canyon and the Weber River at Henefer. This marks the zone of Norwood tuff today.

It appears that a combination of the two theories is closer to the truth than either theory individually. The Wasatch beds were folded into the broad Henefer anticline and then greatly dissected before the Norwood tuff was laid down (Pl. 1, sec. F–F′). A valley as deep or deeper than the present Little East Canyon and Dixie Hollow existed when deposition of the tuff began. On the other hand cross sections C–C′ and D–D′ show both the Tertiary conglomerates and Norwood tuff in Morgan Valley downfolded together, and that the folding chiefly postdated the deposition of the tuff. That the tuff was deposited on a surface of some relief in the Morgan Valley area, however, is indicated apparently by the fact that the conglomerate is thinner on the west flank of the syncline than on the east, and that the Norwood tuff has dips that are divergent in places from the underlying conglomerate. The writer concludes, therefore, that the Henefer anticline was formed mostly in pre-Norwood time, while the adjacent, subparallel Morgan Valley syncline was depressed only slightly. In post-Norwood time the Morgan Valley syncline and the Henefer anticline were accentuated. Durst Mountain became the core of a northward-extending prong of the Henefer anticline as the Morgan Valley syncline was formed and as the frontal Wasatch anticline rose.

The present Great Basin west of the Wasatch was also the site of wide valleys in lower Oligocene time, in which widespread deposition of tuff, marl, clay, and sand occurred. The present valleys and mountains there cannot be taken as an index of past relief because broad folding in post-Norwood time elevated certain parts along the anticlines and depressed others along synclines. In view of this setting, the north-south trending relief feature of Antelope Island and Promontory Range is probably the eroded core of another anticline of this generation, and the valley between it and the frontal Wasatch is probably a broad syncline (Fig. 6).

HERD MOUNTAIN SURFACE

The highest erosion surface, and therefore the oldest surface of which remnants form part of the present landscape in the north-central Wasatch, is conspicuously preserved in the headwater regions of Cottonwood and Lost creeks. The name selected for the surface comes from Herd Mountain, a low, smoothly rounded hill on the high surface, between Cottonwood and Magpie creeks (Pls. 1, 7; Pl. 9, fig. 1). Most of this area is underlain by the nearly level Almy conglomerates, and the high surface at close range seems parallel with the beds. From the peak called Monte Cristo about 25 miles northeast of Huntsville (not in the area of this report), a vast panorama of the surface to the southwest, south, and southeast may be seen. It, also, is mostly underlain by beds of the Wasatch group, but a number of Paleozoic "islands" are beveled by the surface. Its elevation ranges in general from 8500 feet to 9100 feet, but it was not studied with care outside the area of the present report, so the details of slope, and drainage are not well known. Another striking remnant may be seen at the skyline level northeast of Ogden Valley (Pl. 9, fig. 2; Pl. 6, fig. 1).

The remnants of the high surface as seen from the air from Ogden Peak, and from Monte Cristo, do not define a surface with the continuous slope of a pediment such as the highest surface Bradley (1937, p. 174–179) describes on the north flank of the Uintas, but they do not prevent the restoration of a surface with gradual slopes into the sites of the present major valleys. No deposit similar to the Bishop conglomerate (Bradley, 1937, p. 178–179) of the highest surface in the Uintas was seen, but it would not be safe to conclude that such material does not make up some parts of the Herd Mountain surface, because the conglomerates of the Wasatch group underlie much of the area and might not be readily distinguishable from residual or pediment gravels developed from them. The Herd Mountain surface is the higher and older of two prominent surfaces in the north-central Wasatch. In this respect it is similar to the Gilbert Peak surface in the near-by Uintas which is higher and older than the Bear Mountain surface. Moreover, the younger surface in both the north-central Wasatch and the Uintas is a true pediment.

An attempt was made to trace the Gilbert Peak and Bear Mountain surfaces of the Uinta Mountains to the north-central Wasatch by means of the topographic sheets (Marsh Peak, Gilbert Peak, Hayden Peak, and Coalville) and aerial photographs of the Soil Conservation Service than span the two areas. The westernmost remnants of the Gilbert Peak surface on these sheets appear to be the high flat-topped mountains east and west of Smith and Morehouse Canyon and northward across the Weber River Valley opposite the mouth of Smith and Morehouse. Much of the lower and dissected area draining northward into Chalk Creek and westward into the Weber appears to be the Bear Mountain surface. Little field work has been done in this area, but the map evidence seems to warrant the extension of the Uinta names to the north-central Wasatch and to render unnecessary the introduction of new names. It was first planned to do this, but when the attempt was made to correlate the erosion surfaces of southeastern Idaho with those of the north-central Wasatch, marked uncertainty arose, and it was finally considered advisable to use new local names. The surfaces of southeastern Idaho are well known through the detailed work of Mansfield (1927, p. 12–20). A long-standing controversy over the

age of the high surfaces of Idaho and western Wyoming may be followed in the literature, and it is excellently outlined by Mansfield. Before reviewing the problems and evidence, however, it will be necessary to describe the lower surface in the north-central Wasatch and thus to present a fuller picture of the land forms.

WEBER VALLEY SURFACE

The north-central Wasatch area was uplifted after the erosion of the Herd Mountain surface and greatly dissected. The details of the Weber River drainage system, insofar as known, were established during the rejuvenated cycle. In the softer sediments, such as the Cretaceous and Tertiary formations, very wide valleys with extensive pediments as valley sides were eroded. Along the main Weber River Valley the pediment slopes were well developed (Pl. 6, fig. 2; Pl. 11), but, in the drainage area of certain tributaries like Lost Creek, a badland topography is still evident with pediments in various stages of growth (Pl. 10). The pediment is now dissected in most parts of the area, so it may be observed as low ridge crests, flat between subparallel tributaries and gently sloping from the higher mountainous areas to the main valley floors (Pl. 13). The relief model of the area (Pl. 7, fig. 1) shows the pediment less dissected and, hence, more clearly than it actually is because all the minor gullies were not impressed on the model. It gives a valuable picture, however, of the distribution of the pediment and its relation to the high erosion surface. Because the pediment is typically developed along the Weber Valley, it is called the Weber Valley surface.

A profile of the surface across the lower part of Morgan Valley is presented in Figure 5. The line sketch immediately above the profile section is the view beyond looking northward. It shows Ogden Peak, the divide between Morgan Valley and Ogden Valley, and the high erosion surface just west of Herd Mountain. The pediment extends almost to the top of Ogden Peak and has also developed on the other side of the frontal Wasatch south of Weber Canyon to the extent that a dividing ridge is left without conspicuous peaks. The upper sketch of Figure 5 is a detail of the west slope of the frontal Wasatch from Mill Creek to Ford Creek and shows the remnants of the Weber Valley surface. An aerial view of the dissected pediment at Ford Creek is shown in Figure 1 of 13, and of the undissected pediment on Antelope Island in Plate 12. The uniformity of elevation of the crest of the frontal Wasatch may be due to the development of the Herd Mountain surface at that level, but no definite remnants of it exist.

The pediment flanking Morgan Valley and Henefer Valley has a boulder veneer 1–10 feet thick. The boulders are, with rare exceptions, quartzites from the Paleozoic and Proterozoic strata. They were derived from the widely distributed conglomerates of the Wasatch group and, therefore, received most of their rounding in an earlier cycle of transportation. A few volcanic boulders from lenses in the Norwood tuff were noticed on the east slope of Morgan Valley. The position of the boulder veneers is indicated in the profile of Figure 5.

The pediment was broken by block faulting and was either greatly dissected or buried. A large part of the gently dipping surface was buried west of the Wasatch, it is believed, while only the steep, headward reaches of the surface are visible today.

AERIAL MOSAIC OF THE OGDEN CANYON AREA

Ogden Canyon runs across the center of the photograph; Taylor Canyon is the next major canyon on the south; Eyrie Peak is in the northeast corner, and Ogden Peak is in the southeast corner. North is to the left. Heavier white lines are thrust traces, and lighter white lines formational contacts. First thrust near the mouth of Ogden Canyon is Taylor, second is Ogden, and irregular trace at left outlines the capping remnant of Willard thrust sheet. pεf—Farmington Canyon complex, pεpa—Proterozoic strata of the lower argillaceous division, εb—Tintic quartzite, εοε—Ophir shale, €11—lower limestones of Middle (?) Cambrian age, €u—upper limestones of Middle (?) Cambrian age, Dtf—Three Forks (?) limestone, Mm—Madison limestone, Mb—Brazer formation.

SHAPES AND RELATIONS OF THE THRUST SURFACES

(W) Willard thrust surface; (B) Bannock thrust surface; (T) Taylor thrust surface; (O) Ogden thrust surface; (D) Durst thrust surface. The formations of the thrust sheets are not shown. The upward termination of the thrust sheets is approximately their present limits. (1) Pleasant View salient; (2) Eden Pass; (3) Ogden Canyon; (4) Weber Canyon; (5) Farmington Canyon; (6) Eyrie Swell; (7) Huntville sag; (8) Ogden Valley; (9) Morgan Valley; (10) Woodruff Creek; (11) Croyden.

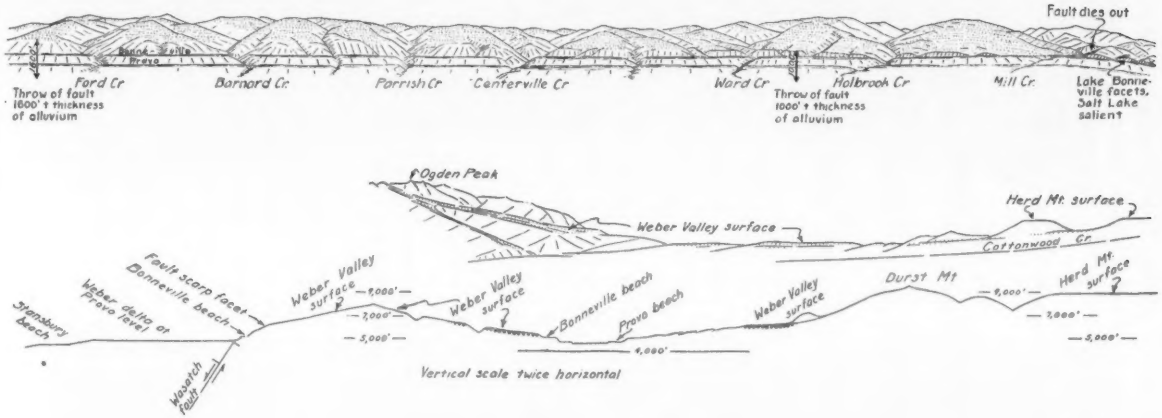


FIGURE 5.—*Herd Mountain and Weber Valley surfaces*

Upper sketch is of the Wasatch front north of the Salt Lake salient. Scarp above Bonneville beach gradually increases in height from Mill Creek to Ford Creek. Pre-faulting surface extends down to the Bonneville Beach at Mill Creek and is about 600 feet above it at Ford Creek. Stippling emphasizes the extent of the Weber Valley surface. Middle sketch shows Weber Valley surface to extend to Ogden Peak on the left but to lie below remnants of the Herd Mountain surface on the right. This sketch is placed in relation to the lower profile as it would be seen in the distance if the observer were to the south of the lower profile looking north.

Gilbert (1928, p. 49–50) recognized and described the "old graded slopes" as "somewhat evenly dressed rib crests . . . . that run from points near the summit of the range to the tops of the frontal facets," and "a definite change of slope from crest line to frontal facet" is conspicuous. These relations are indicated in Gilbert's Figure 37, in which the slopes of the facets are about 30° and the "high-graded" slopes 11°–15°.

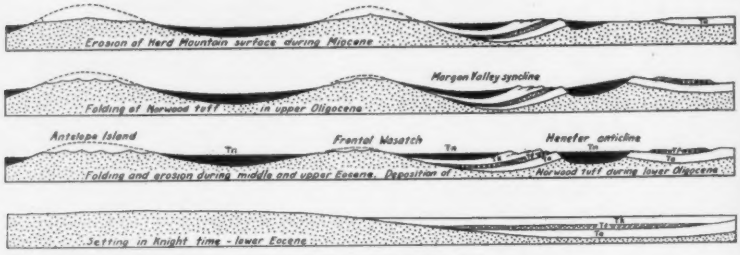

Erosion of Herd Mountain surface during Miocene
Morgan Valley syncline
Folding of Norwood tuff in upper Oligocene
Antelope Island
Frontal Wasatch
Heneter anticline
Folding and erosion during middle and upper Eocene. Deposition of Norwood tuff during lower Oligocene
Setting in Knight time - lower Eocene

FIGURE 6.—*Origin of the Herd Mountain surface*

Sections are approximately to scale but idealized to the extent that they are composites of sections D–D′–D″ and F–F′ of Plate 2. Ta, Almy conglomerate; Tf, Fowkes formation; Tk, Knight formation; Tn, Norwood tuff.

The details in profile of three of the crests and facets between Mill Creek and Ford Creek are presented in Figure 7 of this report. The ridge crest north of Ford Creek slopes about 16°, the crest north of Centerville Creek 11°, and the crest north of Mill Creek 7½°. These "graded slopes" are cause for considerable concern by Gilbert because they are steeper than planed rock slopes in arid regions that he had previously seen, and also steeper than Lawson (1915, p. 34–38) had described in his noted paper on epigene profiles by deserts. He believes, however, that such a result could be attained under exceptional conditions—"for example, those afforded by a small closed basin that was being rapidly aggraded by the waste from a large upland." Bradley (1937, p. 174) measured maximum slopes of only 400 feet per mile near by on the north slope of the Uinta Mountains. A recent review of the Davis and Penck concepts of the retreat of slopes in arid climates, presented by Bryan (1940, p. 260–266), concludes that slopes of hills once formed persist in their inclination as they retreat and that they disappear only when all the volume of rock above the encroaching foot-slopes or pediments has been consumed. Even here the concept of a gently sloping pediment breaking sharply at its upper margin with a much steeper persisting slope does not recognize, seemingly, the existence of pediment slopes up to 16°.

The steep slopes of the ridge crests of the frontal Wasatch offered such a serious obstacle to Gilbert that he chose to think of them as "vestiges of an old peneplain, originally of gentle slope, which acquired its present altitude when uplifted. This interpretation implies that the Wasatch block, or at least a minor block of the Wasatch uplift, was tilted westward as it rose." His general thesis is that the Wasatch is a horst. If so, and if tilting of the horst was westward, the eastern boundary fault would have much greater displacement than the western. Gilbert's respect for bedrock lithology and structure is everywhere evident in his writings on physiographic forms, and had he possessed the advantage of a detailed geologic map

FIGURE 1. NORTH FACE OF OGDEN CANYON

pεf—Farmington Canyon complex, €t—Tintic quartzite, €o—Ophir shale, €l1—lower limestones of the Cambrian, €ul—upper limestones of the Cambrian, Dt—Three Forks (?) formation, Mm—Madison limestone, Mb—Brazer formation, pεp—Proterozoic quartzites, phyllites, and arkosites of Willard thrust sheet. Tintic quartzite repeated by Taylor thrust, T, and Ogden thrust, O. Z-fold barely discernible in Mb just down from Pineview dam. Remnant of high erosion surface in distant mountain beyond Ogden Valley at right, above the letter W. Great Salt Lake and Promontory Range dimly visible at horizon on left. Eyrie Peak directly under symbol pεp on left-center horizon.

FIGURE 2. AERIAL OBLIQUE OF MORGAN VALLEY

View approximately northward. Dissected pediment, inlaid lake beds of the present valley bottom, and hogback to right are important features. Ogden Peak in left background with Weber Canyon at base. Durst Mountain at right. Durst thrust runs along valley between hogback and Durst Mountain.

PANORAMAS OF OGDEN CANYON AND MORGAN VALLEY

FIGURE 1. PRESENT TOPOGRAPHY

Model has same south, west, and north boundaries as geological map (Pl. 1), but extends eastward only to Henefer Valley. (1) Henefer Valley; (2) Round Valley; (3) Morgan Valley; (4) Ogden Valley; (5) East Canyon; (6) Lost Creek Canyon; (7) Cottonwood Canyon; (8) Weber delta; (9) Great Salt Lake; (10) Salt Lake salient; (11) Herd Mountain. Vertical scale about twice horizontal.

FIGURE 2. PRE-FAULTING TOPOGRAPHY

Duplicate model of that in Figure 1 but with clay added to restore the pediment surface. In areas of alluviation (Great Salt Lake Valley and Ogden Valley) layer of clay brings surface to its former position as if by fault movement, not by restoring material that has been eroded away.

when formulating his views of Basin and Range faulting in the Wasatch probably he would have discarded the concept of a major fault in Morgan Valley. No fault is shown on Plate 1, but it has been explained that one of very little displacement may run along the eastern foothills of Morgan Valley. In only a very small amount could it have effected rotation of the frontal Wasatch block westward.

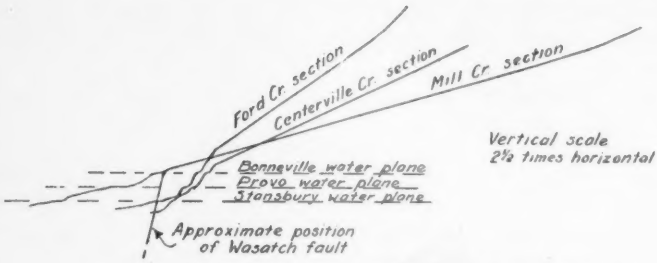


FIGURE 7.—*Profiles of the west front of the Wasatch Mountains*

Profiles labeled by creek names are along ridge crests adjacent to indicated creeks, and section of frontal Wasatch represented is from Mill Creek to Ford Creek.

Although Gilbert saw graded plains and slopes in the "back valleys" of the Wasatch he did not relate them to the dissected surface of the frontal Wasatch. The aerial views, sketches, and map already referred to, together with innumerable observations in the field, all impress the writer very strongly with the identity of the surfaces that slope westward from the frontal Wasatch crest and those that slope eastward; and thus the conclusion that the "evenly dressed" ridge crests of the frontal Wasatch are remnants of the Weber Valley surface. If this is true, the very steep upper slopes were formed at their present inclination and were not appreciably tilted after erosion. The writer does not hesitate to record that such steeply inclined rock-beveled surfaces sloping away from the frontal Wasatch crest exist and that they pass continuously into alluvium-veneered rock plains that are typical pediments. The subject is, perhaps, worth special study and, therefore, should not be extended further here.

AGE OF THE EROSION SURFACES

The assumption has been made in the foregoing pages that the Herd Mountain surface was eroded after the Norwood tuff was deposited and folded. An upper Oligocene or Miocene age of the surface would thus be indicated because the tuff is of lower Oligocene age. The surface could not be much younger than Miocene because sufficient time must be provided for the erosion of the Weber Valley surface, which followed as a result of rejuvenation. The Weber Valley surface had formed before the onset of block faulting, presumably in very late Pliocene or early Pleistocene time. The spacing of these erosion cycles is indicated in the diagram of Figure 2. The assignment of Miocene and Pliocene ages to the Herd Mountain and Weber Valley surfaces, respectively, agrees very well with Bradley's for the Gilbert Peak and Bear Mountain surfaces in the Uinta Mountains with which they seem to

correlate. However, if Mansfield were to extend his observations southward into the north-central Wasatch he would find the same relations there that led him to conclude that his high surface (the Snowdrift peneplain) antedates the Norwood tuff (his Salt Lake formation) and is therefore of Eocene age. This opens the door to a controversy of long standing which would not be entered into if it were not believed that a contribution of value can be made from observations in the north-central Wasatch. Mansfield (1927, p. 354–359) summarizes the controversy and the evidence upon which the conflicting conclusions stand, and only an outline is necessary here.

The guide for dating in Idaho, western Montana, and western Wyoming is the lacustrine and fluviatile beds that in places have yielded Miocene fossils. These beds lie in depressions considerably below the high erosion surface, and nowhere are they known to extend up to the high surface or to be beveled by it. Umpleby believes that the basins in which the beds lie are erosional depressions and that they were eroded after the high surface was established. Pardee believes the old "peneplain" was gently folded across the drainage lines and depressions were thus formed in which the lake beds accumulated (Mansfield, 1927, p. 355–356, Figs. 40, 41). In this sense the high surface runs under the lake beds; in the sense of Umpleby the surface breaks abruptly at high elevations to steep slopes that descend to and under the Miocene lake beds; according to both theories the surface is older than the lake beds.

Blackwelder suggests two alternative hypotheses: (1) The easily eroded lake beds were downfolded or downfaulted between masses of more resistant rock and after a first cycle of peneplanation were subsequently eroded to lowlands. (2) The broad basins occupied by the beds were excavated and filled before the "peneplain" was made. Under both hypotheses the "peneplain" would be younger than the beds. The lacustrine and fluviatile beds are folded and tilted in many basins much more than compaction could account for, and this feature, more than any other, led Blackwelder to favor the first hypothesis. He finds it difficult to see how an Eocene "peneplain" could retain an approximately horizontal attitude and nearly uniform elevation over a large part of two or more States while sediments that have been subsequently deposited upon it were tilted, folded, and faulted to the extent observed.

In the north-central Wasatch the Norwood tuff is the Tertiary formation that occupies the depressions below the high surface, and it does not reach elevations greater than 7000 feet. It is of lower Oligocene age, however, and not Miocene.

The amplitude of the post-Norwood folds—*i.e.*, the frontal Wasatch anticline, the Morgan Valley syncline, and the Durst Mountain anticline—is about 5000 feet. If the Herd Mountain surface were in existence at this time it would have been folded a like amount in the sites of the anticlines and syncline. The absence of remnants of the surface along the frontal Wasatch might indicate that it actually was folded and subsequently eroded in places, but this inference does not hold because the surface is clearly and extensively developed in the site of the syncline between Durst Mountain (anticline) and Lost Creek (Henefer anticline). The syncline is of the post-Norwood generation.

A series of cross sections showing the evolution of the Herd Mountain surface

SALT LAKE SALIENT

View due northward with Salt Lake City in foreground, Salt Lake salient in center, and north-central Wasatch Mountains on horizon. Warm Springs fault scarp, beaches of Lake Bonneville, and Weber Valley surface at top of salient are important features. City Creek drains right side of salient, and ledges along its valley walls are conglomerates of Knight formation of Wasatch group.
(Spence Air Photos.)

FIGURE 1. AERIAL OBLIQUE OF EYRIE PEAK AND WASATCH FRONT
View southeast. (1) Eyrie Peak; (2) Durst Mountain; (3) Uinta Mountains; (4) Herd Mountain. Courtesy of R. E. Marsell, University of Utah.

FIGURE 2. AERIAL OBLIQUE OF OGDEN VALLEY AND MOUNTAINS BEYOND
View northeast. Conspicuous flat-topped mountain in center is composed of folded and truncated Proterozoic quartzites, and high surface is the Herd Mountain. Concave surface sloping from flat top to valley alluvium is Weber Valley surface.

HIGH HERD MOUNTAIN SURFACE

was constructed to test the writer's views of the post-Norwood age of the surface (Fig. 6). The conglomerates of the Almy and Knight formations were derived principally if not wholly from the western highland and were spread extensively eastward. The volcanic ash and tuff of the Fowkes formation appears to occupy a wide depression, and hence a slight amount of folding or erosion may have occurred after the deposition of the Almy and before the deposition of the Knight. At least an episode of volcanism intervened.

Gentle folding and erosion then occurred, and the Henefer anticline became the site of a deep valley because the rocks there were chiefly the softer shales, tuffs, sandstones, and conglomerates of the Henefer formation. The anticlines of the frontal Wasatch and Antelope Island remained high because they are resistant rocks of the crystalline complex. The Morgan Valley syncline and the synclines east and west of the Antelope Island anticline were sites of major valleys. Evidently cross drainage through water gaps connecting the anticlinal and synclinal valleys existed then much as now. The erosion occurred before early Oligocene because the Norwood tuff of that age was deposited in the valleys. The surface upon which the tuff was deposited would be Mansfield's Tygee surface (Fig. 2).

Since erosion has continued to the present on the Norwood tuff in the Henefer anticline and the Morgan Valley syncline, the tuff was much thicker and more extensive at the time of accumulation than now and may have almost drowned some of the ridges between valleys (Second diagram in Figure 6).

Recurrence of folding along the same sites as before—*viz.*, Antelope Island anticline, frontal Wasatch anticline, and Henefer anticline—gently folded the tuff. Here and there in the Morgan Valley syncline rather steep local dips developed presumably due to buckling, while in the Henefer anticline the entire deposit was tilted as shown in the third drawing of Figure 6. This is suggested by the observed dips in the deposit—all to the west.

After the second episode of folding extensive erosion established the Herd Mountain surface. The Norwood tuff was unconsolidated and easily eroded. The Antelope and frontal Wasatch ridges were elevated by the folding, and any tuff that lapped on their flanks was probably removed. If the tuff in the synclines was depressed too much it probably became buried by Miocene sediments. Since Quaternary alluvium masks most of the Tertiary deposits in the basins west of the Wasatch Miocene deposits may occur there in places and have not yet been discovered.

The tuff in the Henefer anticline, upon tilting, was elevated and brought into a position of vigorous erosion, and a large part was probably removed (Bottom section in Figure 6). The Herd Mountain surface as thus established was probably one of extensive pediments from fairly subdued divides, and in the areas of nearly horizontal Wasatch conglomerate the erosion surface was conspicuously gentle in slope.

Upon regional rejuvenation and the establishment of the lower pediment—the Weber Valley surface—the gentle slopes in the areas of nearly horizontal Wasatch conglomerate were left as conspicuous remnants of the high surface. The frontal Wasatch retains no flat-topped remnants because none were there, only the moderately sloping pediments on each side of the subdued divide, and these appear to have been removed in the Weber Valley cycle.

The drainage pattern established consequent to the first episode of post-Wasatch folding is seen in the light of the preceding morphology to have persisted in major outline through the Weber Valley cycle of erosion. The Basin and Range faults did not seriously disturb this pattern—it has persisted until today.

BASIN AND RANGE FAULTING

Wasatch front.—The Wasatch fault is well known through the writings of Gilbert (1928, p. 10–67) and Davis (1903, p. 129–174). Both writers visualized a main fault that hugged the western base of the Wasatch Range and was so recent that the original fault scarp in various stages of dissection still remains. These observations are especially applicable to the Wasatch south of Weber Canyon which is made up of rocks of the old crystalline complex. The top drawing of Figure 5, as previously explained, is a detail of the west front from Mill Creek north to Ford Creek. The stippling represents the remnants of the Weber Valley surface and shows them breaking sharply into a steeper slope that is believed to be the fault scarp. Figure 1 of Plate 13 shows the Weber Valley surface, the fault scarp, and the beaches of Lake Bonneville which are etched on the scarp. The topographic unconformity between the Weber Valley surface and the scarp is shown in profile in Figure 7. The relation of the scarp to the throw of the fault and to the beaches is also illustrated.

Gilbert (1928, p. 16–19) found what he considered direct evidence of the Wasatch fault at the Ogden Reservoir locality (Pl. 5), between Taylor Canyon and Ogden Canyon, and also at Jump Off Canyon just north of Ogden Canyon, in the form of shear zones and slickensides in the hanging wall. The slickensided slopes dip 35°–45°.

The elevation of the sloping pediment, where the fault cuts it, and the amount of movement of the upthrown and downthrown blocks relative to the datum plane determine the position of a water line on the scarp. At the mouth of Mill Creek Canyon the scarp is very small, yet it is so high that the beaches are on the downthrown block. If it were not for a very heavy beach deposit here, bedrock would crop out on the downthrown side of the fault. This aspect of the Salt Lake salient (Fig. 8) will be considered later. At Centerville Creek Canyon the fault has sufficient throw to bring the scarp into the range of the lake-level fluctuations, and the three main beaches are all carved in the upthrown block on the fault scarp. The same is true at Ford Creek, but the scarp is higher. The top of the scarp is slightly above the Bonneville beach at Holbrook Creek (*see* Figure 7) and increases in height very gently to the north but does not exceed 600 feet above the highest beach at any point. The valley alluvium breaks more abruptly with the scarp northward and is much lower in elevation. The scarp, as thus interpreted, has an exposed height of 1600 feet at Ford Creek but gradually decreases southward until it dies out just beyond Mill Creek.

The Wasatch fault cuts higher into the steeply sloping pediment north of Ford Creek, and the resulting dissection has been so extensive just south of Weber Canyon on the west front that no clearly recognizable remnant of the pediment remains. It is therefore difficult to determine at what elevation the top of the scarp once was (Pl. 7). Triangular facets may still exist after considerable erosion of the fault scarp, but the top of the facets will not mark the top of the original scarp (Davis, 1903, p. 129–174).

The height of the scarp exposed does not mark the throw of the fault if the downthrown block is covered with alluvium. The thickness of the alluvium at the fault must be added to the height of the scarp. The thickness of the alluvium is not known except at the sites of a few deep wells located some distance from the mountain front. Even in these wells the existence of the Norwood tuff (Salt Lake formation, Eardley and Haas, 1936, p. 69–76) has only recently been recognized, and the contact with the overlying alluvium is a matter of doubt. In a well west of Brigham City the alluvium was 500 feet deep, and a well northwest of Salt Lake City was drilled through volcanic tuff at 1302 feet and had probably penetrated the Norwood tuff proper at a much shallower depth. At many places around the north end of Bear River Bay and Great Salt Lake, west of the Promontory Range, tuffs, clays, and marls crop out without any alluvial covering. It thus appears that the alluvium at the mountain front is not more than 1000 feet thick, even in the sites of pre-existing valleys now buried, and thins to nothing at the Salt Lake salient (Fig. 8). The throw at Ford Creek would be, therefore, 1000 feet (alluvium) plus 1600 feet (height of scarp) which equals 2600 feet. The height of the scarp increases to about 2000 feet just north and south of Ogden Canyon (if certain facets there are fault scarp facets), and, therefore, the throw at this locality is of the magnitude of 3000 feet.

It is evident that the writer does not follow Davis (1903) in postulating block faulting as the cause of the entire relief of the range. Davis visualized a peneplain before faulting, a vertical displacement along the fault of approximately 10,000 feet, and almost all the drainage pattern as a development subsequent to faulting. The results of the present study demonstrate that the mountains of the north-central Wasatch were approximately as high before faulting as after, that the fault has a maximum displacement of about 3000 feet at Ogden and Weber canyons, but decreases gradually southward until it dies out completely at the Salt Lake salient, and that the entire drainage pattern predates the faulting, except in the areas of alluviation. These conclusions parallel approximately earlier ones that the writer arrived at in a reconnaissance physiographic study of the whole range (Eardley, 1933, p. 243–267). Gilbert's (1928) extended study of the drainage of the Wasatch will be discussed later.

Salt Lake salient.—A Basin and Range fault extends from Warm Springs on the north side of Salt Lake City to and beyond Beck's Hot Springs (Fig. 8; Pl. 8). A typical fault scarp along this section is the principal evidence of the fault's existence and position. It bounds the west side of the Salt Lake salient and has been pictured by Gilbert (1890, Pl. 44; 1928, Pl. 8A). He identifies small hanging valleys in the upthrown block and an alluvial fan at the base of the scarp cut near the apex by recent movement on the fault. Pack (1926, p. 399–410) also described this section of the "Wasatch fault," and Schneider (1925, Fig. 2) pictured a slickensided face uncovered in a gravel pit near Warm Springs. The local slickensided face dips 72° W., and the slickensides indicate dip-slip movement. The nature of the fault and the development of the Salt Lake salient as postulated by the writer is illustrated in Figure 8. The lower diagram shows the topographic expression of the salient and adjacent Wasatch Mountains as they are today. The scarp that terminates the Weber Valley surface and that dies out at Mill Creek has already been described. This is properly the Wasatch fault. The front of the salient from Mill Creek to a

point 2 or 3 miles north of Beck's Hot Springs does not display characteristics of a fault scarp because the Weber Valley surface extends unbroken except for beach cliffs down to and under the alluvium. Therefore, no fault is believed to exist along

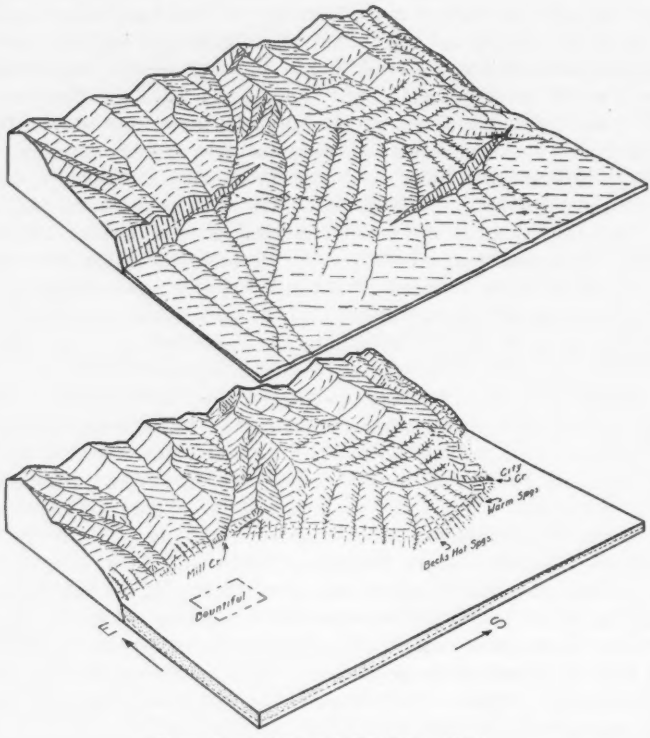


FIGURE 8.—*Origin of the Salt Lake salient*

Topography is generalized from aerial photographs. Upper diagram shows pre-faulting topography and uneroded fault scarps. Dashed line is elevation to which alluvium will accumulate after faulting and is shown in the lower diagram. Lower diagram shows region as it is today. Compare with Plate 8.

this face. The scarp that bounds the west face of the salient is prominent and clear in its topographic expression. It is little dissected and conveys the impression of youthfulness. The steep, fairly straight fault scarp contrasts conspicuously with the gently sloping, somewhat irregular north and south faces of the salient. By postulating faults only where true fault-scarp features can be detected, by removing the alluvium that accumulated after faulting, and by restoring the material eroded from the upthrown blocks, the pre-faulting topography has been visualized and diagrammed in the upper drawing of Figure 8. The dashed line indicates the upper limit of accumulation of alluvium today, and the lower limit of dissection of the fault scarps. This theory does not necessitate bending a branch of the Wasatch fault inordinately westward around the salient, as Gilbert does (1928, Fig. 12), and

FIGURE 1. VERTICAL AERIAL PHOTOGRAPH OF PINE CANYON
Lost Creek Valley in lower right corner. Tributaries extending to upper left corner have about established pediment, but other tributaries are still in badland stage. Rocks being eroded are nearly flat-lying beds of Almy conglomerate.

FIGURE 2. AERIAL OBLIQUE OF LOST CREEK VALLEY
Badland topography most conspicuous, but overall sloping ridges toward valley and approaching pediment stage of erosion is discernible. Herd Mountain surface at horizon. Badlands and partial pediment constitute here the Weber Valley surface.

FIGURE 1. AERIAL OBLIQUE OF WEBER VALLEY DOWNSTREAM FROM ECHO RESERVOIR
Dissected pediment flanks valley. Echo Canyon with Pulpit conglomerate cliffs just beyond dam. Durst Mountain in the distance.

FIGURE 2. AERIAL OBLIQUE UP LITTLE EAST CANYON FROM THE WEBER
Pediment remnants conspicuous in lower part of valley, and in divide area (middle distance) pediment is still undissected.

PEDIMENT IN WEBER VALLEY

at the same time it embodies the concept of an extensive basinward-sloping Weber Valley surface that existed before faulting. Gilbert needed also to extend the main Wasatch fault across the salient to make his theory of the origin of the salient tenable. Schneider (personal communication) and the writer have found no fault there.

On the basis of the theory of origin of the Salt Lake salient thus presented, and confirmed in general by his studies on the Traverse Range south of Salt Lake City R. E. Marsell (personal communication) located a water well for the Water Department of Salt Lake City in the mouth of City Creek Canyon. The site at Fourth Avenue and Canyon Road would be where the pediment is overlapped by alluvium according to the present theory, but where the alluvium should be very thick on the downthrown side of the Wasatch fault, according to Gilbert. The predicted depth to bedrock, on the theory that no fault exists, was 500 feet, and the depth at which the drill entered bedrock (the Knight formation) was 465 feet.

A small indentation of the Wasatch front exists at the mouth of Weber Canyon, and in Figure 9 the origin of such a feature in the light of strong relief before faulting is indicated. It is concluded that the aspect of the re-entrant depends on: (1) the position of the fault in relation to the mountains and flanking pediments; (2) the throw of the fault; (3) the elevation to which alluvium accumulates; and (4) the extent of dissection.

Weber Canyon, according to this concept, was the site of a major transverse valley at the Weber Valley surface stage and is believed to have been not unlike the valley pictured in the top diagram of Figure 9. The present aspect differs, however, from that conveyed in the lower diagram in that the alluvium has not yet buried the scarp in the central valley area. Shoulders are present, admittedly somewhat inconclusive, that seem to betray the extent of the original scarp (Pl. 7; Pl. 13, fig. 2). Gilbert's Plate 25B of the mouth of Weber Canyon shows a smaller, relatively mature canyon on the right and an alluvium-capped bench that extends several hundred feet up the spur above the Bonneville beach. These features apparently define the pre-faulting surface and indicate the amount of entrenchment in post-faulting time at the mouth of the canyon.

Pleasant View salient.—An impression of the shape, size, and relation of the Pleasant View salient to the main range may be obtained from Figure 10A. Gilbert's photograph (1928, Pl. 9A) is especially good. The areas of bedrock outcrop on it are shown on Plate 2. The Wasatch fault is believed to separate the range from the salient, and, according to Gilbert, the evidence for the fault consist[illegible] part in the continuity of the range features from the escarpment back of the s[illegible] to the sheared quartzite at Willard. He also notes a fresh scarp in the valley alluvium which is continuous from Ogden Canyon northward across the salient to Willard with only a short gap where most recent alluvium has washed out of Barett and Wash canyons.

Section A–A′ (Pl. 2) is regarded as the most logical interpretation of the structure and according to it the throw of the Wasatch fault is slightly more than 2000 feet. The maximum throw of the fault south of Ogden Canyon was figured to be 3000 feet and not to exceed 2500 feet for most of its length to the Salt Lake salient, where it dies out. A throw of 2000 to 3000 feet is that used in the diagrams of Figure 10, which show the evolution of the salient.

Does a Basin and Range fault occur along the outer margin of the salient? Gilbert

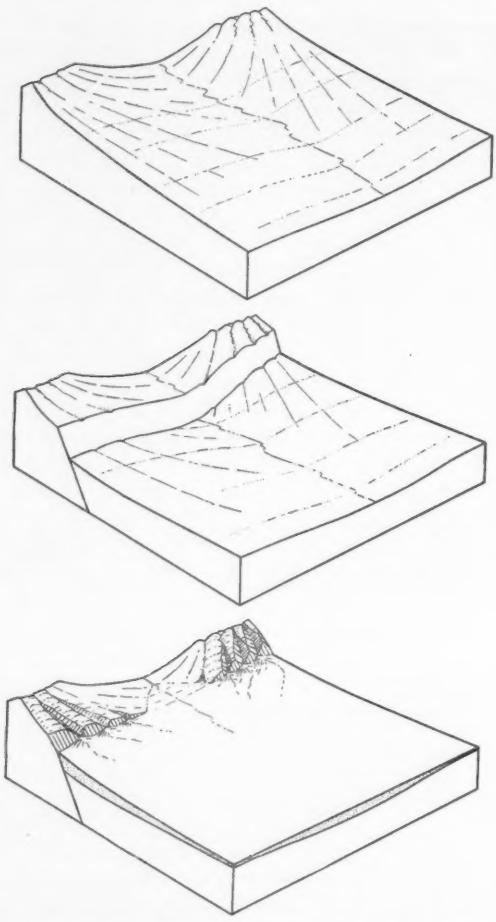


FIGURE 9.—*Development of a re-entrant in a fault scrap*

In the upper diagram the valley cuts through a long north-south-trending range like the north-central Wasatch, and extensive pediments flank the cross-cutting valley and the range. Normal faulting occurs in the second diagram. The fault is not quite parallel with the range and cuts the pediment at a higher elevation on the right than on the left, but the throw is assumed to be uniform. The scarp is in part buried and in part dissected in the third diagram. Alluvium has accumulated so extensively that it has flooded up the valley somewhat even over the upthrown block.

was inclined to think so but made it clear that the evidence is very meager. Bedrock rises abruptly from the alluvium, but the escarpment is irregular in plan view and in slope. Gilbert curved his fault incredibly around the ends of the salient in order to fit the crags that project above the alluvium. Hot sulphur springs are the best evidence for a fault known to the writer.

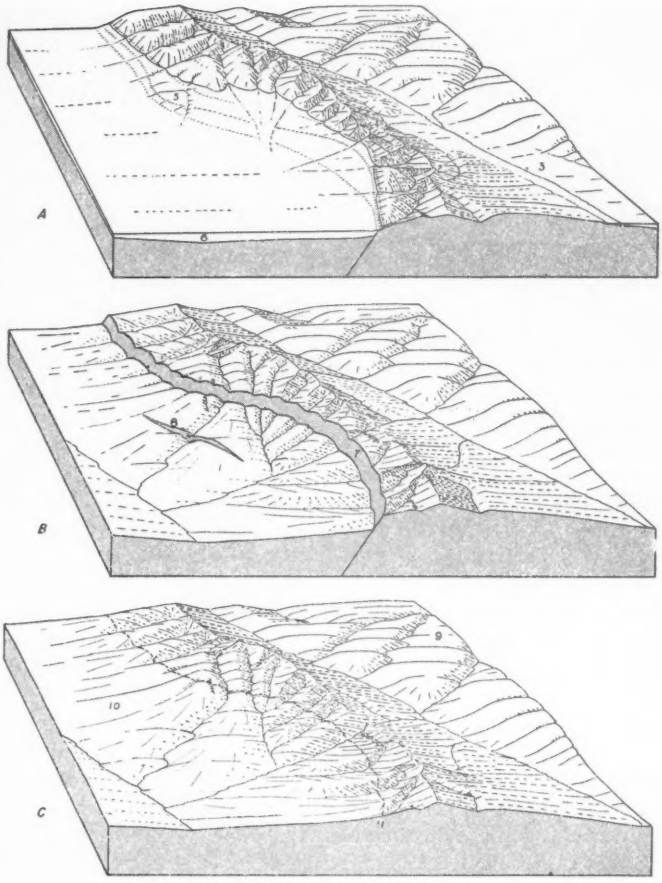


FIGURE 10.—*Origin of the Pleasant View salient*

Upper diagram represents the present surface. 1, Ben Lomond; 2, Eyrie Peak; 3, Ogden Valley; 4, Eden Pass; 5, Pleasant View salient; 6, alluvium. Middle diagram shows Wasatch fault (7) and Pleasant View fault (8) as if no erosion had occurred during the faulting. Lower diagram shows the topography before faulting. 9, Herd Mountain surface; 10, Weber Valley surface (pediment surface); 11, position of Wasatch fault.

If a frontal fault exists, it can be considered as similar to the Warm Springs fault of the Salt Lake salient, but with its scarp buried more by the valley alluvium. The throw of the Pleasant View fault may be very small, and if so the scarp would not require much alluvium to bury it. A very small fault is diagrammed in Figure 10.

Eden Pass is about 5800 feet above sea level, 600 feet above Liberty in Ogden Valley, and 1300 feet above North Ogden in the Great Salt Lake Valley. Gilbert (1928, p. 58, 59) pointed out the following features of the pass (Pl. 2): (1) It is broad

and alluviated; (2) its summit lies close to Ogden Valley with a steep eastward slope and a gentle westward slope; (3) the tributaries of the westward drainage (North Ogden Canyon) are not barbed but arborescent; (4) three shoulders west of the pass suggest the elevation and gradient of the stream before faulting and dissection. From these observations he concluded that the pass was a water gap at one time and that part of the drainage of Ogden Valley flowed through it. He believed that the gap was in existence and was being deepened through the early stages of faulting and that the stream running through it from east to west was captured by the Ogden River. At least 1000 feet of displacement occurred after capture, according to Gilbert. Since the evidence presented in the present report indicates that faulting did not begin until the Weber Valley surface had been formed, the writer believes that the stream through Eden Pass was decapitated during the Weber Valley cycle of erosion. In fact, the summit of the gap is over 600 feet above the Weber Valley surface in Ogden Valley, and, therefore, a similar amount of downcutting after capture occurred before faulting began. It would, therefore, appear that the piracy occurred in the early stages of the Weber Valley cycle. The conclusions are illustrated in Figure 10.

Dissection of the Weber Valley surface.—The pre-faulting pediment is extensively dissected in the Weber Valley east of the Wasatch fault, and the amount of dissection decreases eastward. The evident conclusion is that the controlling element of stream activity in the Weber Valley was the Wasatch fault. The features of dissection and alluviation are illustrated in the models of Plate 7.

Gilbert's theory.—In the foregoing discussion of Basin and Range faulting the posthumous work of Gilbert (1928) was referred to principally where his observations and conclusions supported the writer's. But if the report is carefully studied it will be found that Gilbert has described several local physiographic settings that lie within the area of this report and has built up from these and other parcels of information, in a very skillful fashion, general structural and physiographical conclusions that differ at many turns from the writer's. It does not seem feasible to consider here each of his localities and each of his conclusions. Instead, it is proposed to list the highlights of each theory and especially to point out the features the visiting geologist must examine in the field to convince himself of the relative value of the theories.

(1) Both theories assign an important role to block faulting in the late physiographic development of the range.

(2) Both theories recognize erosional surfaces of some complexity before faulting. Gilbert speaks of a high "grade plane" which most probably is the writer's Herd Mountain surface. He also found lower grade plains at different localities but did not correlate them nor see in them a single extensive dissected pediment as the writer does.

(3) The writer visualizes the mountains almost as high before block faulting as today, because he believes the dissection of the Weber Valley surface was caused by the faulting and that alluviation on the downthrown block about compensated for the elevation of the upthrown block. Gilbert studied the "back valleys" (Morgan

WEBER VALLEY SURFACE ON ANTELOPE ISLAND

Aerial oblique looking northward over Antelope Island of Great Salt Lake. Advanced stage of pedimentation seen from the air is not easily recognized from the ground. Lake Bonneville at its highest level covered all but the highest peaks and softened somewhat the rugged nature of pediment slopes by deposition in gullies. Geology of island shown on Plate 3.

FIGURE 1. WEST FRONT OF WASATCH AT FORD CREEK CANYON
Aerial oblique showing pediment remnants terminating at fault scarp. Lake Bonneville beaches conspicuous but still minor features on triangular facets of scarp. (Courtesy U. S. Forest Service, Ogden, Utah. U. S. Dept. Agric., Misc. Pub. 196, Fig. 3, 1934.)

FIGURE 2. WEBER CANYON THROUGH THE WASATCH
Aerial oblique westward showing V-shaped water gap, and Weber delta in the Great Salt Lake Valley faintly in the distance. Remnants of pediment in middle right foreground indicate that canyon at its upper end was almost as deep before dissection as now. Lower end of canyon believed to have been dissected more than upper end due to tilting during Basin and Range faulting. (*See* Gilbert, 1928, Pl. 25B.)

DISSECTION OF WEBER VALLEY SURFACE

and Ogden valleys), "air gaps" (Eden Pass), cross drainage (Weber and Ogden canyons), and areas of mature aspect, and he reasons that these features must have been in existence when the "Pliocene lake beds" (the Norwood tuff of Oligocene age) were deposited. The deep cross canyons such as Weber Canyon and Ogden Canyon were antecedent to the uplift of the frontal Wasatch, he believed, and set the stage for water-way connections between the back valleys and the Great Salt Lake basin during the deposition of the "Pliocene beds." Careful search of Gilbert's writings reveals that he was noncommittal on the nature of the uplift responsible for the antecedent drainage, but it would be easy to infer that he was thinking of early stages of movement along the Wasatch fault.

In the writer's opinion the drainage pattern was established in outline consequent to and immediately after the post-Knight folding. The deposition of the Norwood tuff and the gentle post-Norwood folding undoubtedly changed some of the details, but the sites of the major depressions persisted. The cycle of erosion following the post-Norwood folding resulted in the Herd Mountain surface and ended when the region was uplifted. The streams are antecedent to the post-Norwood folding but only in small part could they be antecedent to the post-Wasatch folding. At that time the drainage was eastward from the western highland into the area of accumulation of the Wasatch sediments. The major water gaps were, perhaps, present when the Norwood tuff was deposited and had certainly become established as part of the Herd Mountain surface. They were deepened upon regional uplift and dissection of the Weber Valley surface.

(4) Gilbert came to view the frontal Wasatch as a horst between the Wasatch fault and a fault along the back valleys and probably associated the drainage features with the growth of such a structure. Detailed mapping has shown that if a Basin and Range fault occurs along the east side of Morgan Valley it is of very small throw and is a relatively small feature superposed on the broad deep valley already existent. In fact, the writer's concept of the major Wasatch fault is one that assigns it little influence in changing or controlling the drainage pattern. It has imparted to the range in most sections, however, a fairly straight and abruptly rising front from an alluviated piedmont.

LAKE BONNEVILLE TERRACES AND DEPOSITS

The Wasatch front.—The highest shore line of ancient Lake Bonneville is scarred upon the Wasatch front at an elevation of 5200 feet. This is called the Bonneville beach and is about 1000 feet above Great Salt Lake. The position of the Bonneville beach and the next two lower ones of extensive development, the Provo beach and the Stansbury beach, are marked on Plate 1. The recognition, description, and decipherment of these beaches are the work of Gilbert, and for a history of Lake Bonneville his Monograph 1 may be referred to. The principal, yet small, contribution of the present report to this phase of the physiography of the area is the mapping of the three beaches.

A problem related to the Wasatch fault but not previously mentioned is the relation of the wave-cut facets of Lake Bonneville to the facets of the fault scarp. Much

confusion has arisen because of failure to recognize and distinguish the two, or to appreciate the possibility of a prominent row of wave-cut facets in the pediment slopes of the Weber Valley surface in the absence there of a fault scarp.

Gilbert (1928, p. 5) assumed that the slope of the fault scarp was the dip of the fault plane because he found several slickensided surfaces which dip about 35°. This low dip was taken as the inclination of the main Wasatch fault. The problem has been reviewed by Gilluly (1928, p. 1103–1130) for the Oquirrh Range, and he concludes that the scarps there dip at lower angles than the faults. A confirmation of this view is found near Warm Springs where a slickensided surface on the Warm Springs fault dips 70°, and the scarp above dips less than 35°. The slickensided surfaces observed by Gilbert at half a dozen localities along the Wasatch front, all dipping 33°–35°, may be due to landslide shearing along the steep and newly elevated fault scarp. The landslide masses are probably completely concealed in many places by younger alluvial fans and lake beds, but large Bonneville and post-Bonneville slides are known in the southern Wasatch, just north of North Ogden Canyon (Pl. 2), and at Brigham City.

Morgan Valley and the Weber delta.—The lake at the Bonneville level extended as a narrow arm up Weber Canyon and then expanded into Morgan Valley upstream approximately to Morgan (Pl. 1). Considerable deltaic material was deposited in the Morgan embayment at the Bonneville stage, and the upper surface of the deposit slopes with several minor terraces as interruptions from Morgan at the Bonneville level to Peterson at the Provo level. Extensive flat areas at the Provo level below Peterson attest the near or complete filling of the Morgan bay with sand and silt at this time. These sediments and also those of the Bonneville stage were deposited in the valley bottoms that had been freshly incised in the Weber Valley surface as the result of Basin and Range faulting.

The large Weber delta at the mouth of Weber Canyon (Pls. 1, 7, 14) was built at the Provo stage. Only small parts of the delta extend above the Provo level and indicate that not much material was transported through Weber Canyon at the Bonneville stage. The load of the Weber River was dropped in the bottom of the Morgan embayment at this time, and clear waters issued from Weber Canyon. At the Provo stage the lower part of Morgan bay below the Provo level, and also the Weber Canyon, were filled and permitted the Weber to carry its entire load down to the mouth of the canyon and deposit it in the delta.

Ogden Valley and the artesian reservoir.—Ogden Canyon and Ogden Valley provided another bottle-shaped embayment for Lake Bonneville. The waters spread in Ogden Valley fan-fashion from the head of Ogden Canyon and covered about two-thirds of the floor of the wide valley. This is known from the spread of laminated lake-bottom clays rather than from shore features. According to Leggette and Taylor (1937, p. 99–161), who studied the ground-water conditions of Ogden Valley, the lake sediments have a maximum thickness of 60 feet near the head of Ogden Canyon, are impervious, and are the cause of artesian pressure in the area that they cover. The lake clays were deposited on a coarse and porous alluvium that had been spread as alluvial fans from the mountain slopes. The several forks of Ogden River flow over the porous material before reaching the clay, and much of

their water seeps into the alluvium, then moves to the artesian reservoir under the clay. Leggette and Taylor determined inflow by both stream gauging and water-table fluctuation. The outflow is in part artificial and in part natural. The artificial is measured simply by well discharge, and the natural (leakage into Ogden Canyon) by stream gauging. Increasing demands by the city of Ogden on the artesian reservoir had resulted in the application of compressed air to stimulate flow during the summer, and the artesian pressure had decreased alarmingly. The ground water of Ogden Valley is amenable to fairly complete measurement, and the rate of discharge and recharge were brought so well in hand by Leggette and Taylor that they were able to conclude that the present forced rate of artificial withdrawal is within the safe yield of the reservoir.

Prospects of ground water in Morgan Valley.—Undoubtedly the artesian basin in Ogden Valley has led to speculation on the artesian conditions in Morgan Valley, a closely related back valley of the Wasatch. No flowing well water has been produced in Morgan Valley, although at least one shallow well has been drilled. One difference is immediately evident—*viz.*, the Weber Valley surface in Morgan Valley is dissected, and its remnants are 100 to several hundred feet above the valley alluvium, whereas the pediment surface in Ogden Valley extends down to and under the alluvium. It seems as if faulting and tilting of the Wasatch block made Ogden Valley a trap for alluvium and that downcutting of Ogden Canyon did not proceed rapidly enough to keep the wide valley drained. At the same time rejuvenation of the Weber River by faulting resulted in the extension of vigorous erosion through Morgan Valley and many miles farther up the Weber Valley with no opportunity for alluvium to accumulate. An artesian reservoir in alluvium would, therefore, not be expected in Morgan Valley.

This consideration does not exhaust the ground-water possibilities, however. Morgan Valley is eroded into a broad, gentle syncline in the Knight formation and the Norwood tuff. The coarse conglomerates of the Knight are porous in outcrop and undoubtedly are good aquifers. In addition, the water they carry would be confined as a result of the synclinal structure. The Weber River cuts the conglomerates of the west flank of the syncline (Weber Canyon) and probably drains any water in them above the elevation of the cut in its immediate vicinity. A hydrostatic head sufficient to raise water above this level in a well in the syncline would not be expected. A great reservoir of underground water must exist in the syncline, and wells drilled into the conglomerates would probably fill almost to the surface. Raising the water only a few feet by pump would be necessary to supply needs down river. The wells would have to be drilled from 500 to 2000 feet deep.

Ground water in the Great Salt Lake Basin.—A well drilled at the site of the gas regulating station of the Mountain Fuel Supply Company in sec. 29, T. 5 N., R. 1 W., tested over a million gallons of water a day with only a 3-foot drawdown of the water table (J. W. Proctor, personal communication). Another well recently completed by the Command of the Ogden Ordnance Depot in the SE 1/4 NE 1/2 sec. 30 of the same township started at a surface elevation of 4785 feet is reported to have reached a depth of 900 feet. The depth to which the well of the Mountain Fuel Supply Company was drilled is not known nor is a log available, but it is instructive

to know that a flow of about 1½ second feet has been obtained from the well. It has long been known that the alluvium of the Great Salt Lake Basin is an abundant source of water, and considerable study has been made of the ground-water conditions in the basin. Alluvial fans in general are well known as sources of ground water.

Special attention has not been given, however, to a basin filled with alluvial fans where the bedrock floor is one of considerable relief. In Figure 9 the pre-faulting valley of the Weber River is represented as cut and displaced by the Wasatch fault with the burial of the part eroded in the downthrown block. The site of the buried valley is one of much deeper alluvium than elsewhere and would have been filled shortly after the onset of faulting with coarse material. It is also known that the alluvium becomes finer basinward and gives way to clays, peats, and marls. Some natural gas has been found in wells drilled in the fine sediments, but water is scarce in them. The best artesian wells are located several miles from the mountain front down the slopes of the alluvial fans but not far enough basinward to miss the porous gravels. Instead of paralleling the mountain front, as previously believed, the areas of optimum artesian pressures would be over the buried valleys and some distance from the mountain front. This concept helps explain the local artesian areas at Bountiful and Farmington, mapped several years ago but not fully understood at the time (Eardley and Haas, 1936, Fig. 5), and seems to be a guide to the artesian area now being developed opposite the mouth of Weber Canyon by the Command of the Ogden Ordnance Depot.

POST-BONNEVILLE EVENTS

The most conspicuous effects of post-Bonneville erosion are the trenching of the Weber delta in the Great Salt Lake Valley and the lake beds in Morgan Valley by the Weber River. A parallel trenching by the Ogden River of its delta and of the lake beds in Ogden Valley took place at the same time. These features are fully described by Gilbert (1890) "Lake Bonneville," U. S. Geol. Survey, Monograph 1.

The activity of the Weber River during Lake Bonneville time and afterward above the highest point of still water (about at Morgan) was not investigated. Careful attention to the deposits and terraces in the Weber Valley from Morgan to Kamas might yield significant results not only pertaining to the geologic history in post-Bonneville and Bonneville time, but to immediate pre-Bonneville time.

Movement along the Wasatch fault in post-Bonneville time has been recognized by Gilbert (1890, p. 340–352) and expertly illustrated and described. He concludes as follows: A fresh scarp cuts the apex of an alluvial fan north of Warm Springs (Gilbert's Pl. 44) on the fault here called the Warm Springs fault. The throw is a composite of three individual movements of 15, 5, and 10 feet each, making a total of 30. The fresh scarp has not been washed by Lake Bonneville waters and is, therefore, younger than the lake at the Stansbury level, below which it occurs. Another fresh scarp in the alluvium runs along the Wasatch front continuously from Mill Creek (?) to North Ogden Canyon. Gilbert's Figure 46 shows the scarplet at the mouth of Farmington Canyon, and his Figure 47 is a profile across fresh scarps near Ogden. The throw ranges from 25 to 75 feet (it offsets the Weber delta 50 feet) but is a composite in places of several scarps where blocks of alluvium are dropped

FIGURE 1. WASATCH FRONT BETWEEN FARMINGTON AND KAYSVILLE
Vertical aerial photograph of Shepard Creek and Bears Creek (north) showing three beaches from highest to lowest—Bonneville, Provo, and Stansbury.

FIGURE 2. PART OF WEBER DELTA
Vertical aerial photograph showing modern Weber Valley (top) cut in Weber delta (bottom). Prevailing down-canyon winds have furrowed delta sand. Scale of photographs about 1¼ inch to the mile.

as small graben and also tilted. The vertical aerial photographs of the north-central Wasatch piedmont do not reveal these scarplets with any certainty.

An almost constant wind that issues into Salt Lake Valley from the mouth of Weber Canyon, and which at times attains near-gale velocities, has been a competent agent of transportation in post-Bonneville time. The fine sands of the top-set beds of the Weber delta have been shaped by the Weber Canyon wind into gently rounded longitudinal dune ridges and furrows, as may be seen in Plate 14. It is possible that the removal of sand has resulted in the shallow furrows, and that the ridges between are not accumulations of sand but erosional remnants. This aspect was not studied in the field. It is known only from the aerial photographs.

CONCLUSIONS

New formation names in the north-central Wasatch, deemed necessary as a result of the present study, are the Henefer formation (lower Paleocene or uppermost Montana), the Pulpit conglomerate and Saw Mill conglomerate as divisions of the Almy conglomerate (middle Paleocene), and the Norwood tuff of the Salt Lake group (lower Oligocene).

The stages of deformation have been outlined in Figure 2 and summarized in the abstract, and for that reason they will not be further reviewed. The structural pattern and its origin may be summarized as follows:

Three thrusts, the Taylor, Ogden, and Durst, strike northwestward through the north-central Wasatch and dip northeastward. The three thrusts are truncated by the Willard thrust, which in general also strikes northwestward and dips northeastward. Available evidence favors the hypothesis that the three lower thrust sheets were thrust westward against a small buttress of Archeozoic rocks, but the movement of the Willard sheet is not clear. The great Bannock thrust is traced southward to a position opposite the Willard where the two thrusts dip toward each other. Two hypotheses of the origin of the thrusts are reviewed. One depicts the complimentary relation as an orogen in which the Willard moved westward and the Bannock eastward. The other visualizes the relation as a single great sheet that is rooted in the Fremont Proterozoic trough over 20 miles west of the trace in the Wasatch Mountains. Both hypotheses have weak points in light of existing information, but the one of eastward thrusting has sufficient merit not only to justify but to require an attempt at tracing its extent southward. A possible route of the trace is one that connects the Willard with the Sheeprock thrust of west-central Utah and this with thrusts in southern Nevada. A thrust sheet is thus postulated which extends at least from the Snake River through southeastern Idaho into northern Utah, and through central and southwestern Utah into southern Nevada, a distance of nearly 700 miles. The sheet would have a breadth between salients and embayments in its front of 45 miles.

The Taylor, Ogden, and Durst thrust sheets are strongly folded normal to their general strike. These cross folds are parallel with and of the same age as the east-west folds in the central Wasatch. Together they are probably continuations of compressional structures in the Idaho-Wyoming arc which veer westward at the approach to the Uinta arch. The buttress against which the sheets were thrust was

itself folded by the cross folds. Considerable erosion followed, and then the great Willard sheet was thrust westward or eastward over the beveled edges of the older sheets. Cross folding after the Willard thrusting represents a recurrence of compressional deformation in this area transverse to the direction of general thrusting.

Three orogenic deposits are found in the area—the Kelvin, the Henefer, and the Wasatch. The Kelvin predates any thrusting displayed in the Wasatch and may indicate a Lower Cretaceous stage of thrusting in the western highland. The Henefer is a result of either the Taylor, Ogden, and Durst thrusting and cross folding or the early stages of Willard and Bannock thrusting. It may be overridden in part by the hypothetical Willard-Bannock sheet. The Wasatch conglomerates come from an uplifted western highland and partly bury the extensively eroded Willard sheet.

In the closing stages of the Laramide revolution gentle north-south folding in the area of the highland and in the Morgan Valley and Ogden Valley area was superposed upon the older eastward- and northeastward-trending folds and the northwest-striking thrust sheets. The Henefer anticline was accentuated in post-early Eocene time, and the Morgan Valley syncline was formed mostly in post-early Oligocene time. The main frontal Wasatch consisting chiefly of the crystalline complex became the core of a gentle anticline parallel with the Morgan Valley syncline, and the Durst Mountain "island" became the core approximately of another Tertiary anticline on the east.

Following the gentle north-south folding rocks of all ages were bevelled to an extensive surface with relief perhaps not more than 1000 feet—the Herd Mountain surface. Regional uplift followed the Herd Mountain erosion cycle, and the easily eroded Norwood tuff became the site of large valleys both east and west of the frontal Wasatch. The drainage pattern in its broader features was established in pre-Norwood time, and most of the details were impressed during the Herd Mountain erosion cycle. Upon uplift they became altered only in local details. During and following the uplift wide pediment slopes to the master stream of the region, the Weber River, were eroded and are called the Weber Valley surface. The major aspects of the present topography were established by this time.

Basin and Range faulting interrupted the pediment cycle of erosion and resulted generally in dissection of the pediment in the upthrown blocks and burial by alluvium of the pediment in the downthrown blocks. The Wasatch fault has a maximum displacement of about 3000 feet at Ogden and Weber canyons and dies out southward at the Salt Lake salient. The zone of disturbance is continued southward, however, by the Warm Springs fault which parallels the Wasatch fault but is offset westward several miles from it. The Warm Springs fault scarp marks the western limit of the Salt Lake salient, a spur of the Wasatch Range already existent before faulting began.

The local relief after faulting was only slightly greater than the pre-faulting relief.

REFERENCES CITED

Baker, A. A., and Williams, J. S. (1940) *Permian in parts of Rocky Mountain and Colorado Plateau regions*, Am. Assoc. Petrol. Geol., Bull., vol. 24, p. 617–635.

Billings, Marland (1933) *Thrusting younger strata over older*, Am. Jour. Sci., 5th ser., vol. 25, p. 140–165.

Blackwelder, Eliot (1910a) *New light on the geology of the Wasatch Mountains, Utah*, Geol. Soc. Am., Bull., vol. 21, p. 517–542.

——— (1910b) *Phosphate deposits east of Ogden, Utah*, U. S. Geol. Survey, Bull. 430, p. 539–551.

Boutwell, J. M. (1907) *Stratigraphy and structure of the Park City mining district, Utah*, Jour. Geol. vol. 15, p. 434–458.

———, and Woolsey, L. H. (1912) *Geology and ore deposits of the Park City mining district, Utah*, U. S. Geol. Survey, Prof. Paper 77, 231 pages.

Bradley, W. H. (1936) *Geomorphology of the north flank of the Uinta Mountains*, U. S. Geol. Survey, Prof. Paper 185 I, p. 163–199.

Branson, E. B. (1927) *Triassic-Jurassic "Red beds" of the Rocky Mountain Region*, Jour. Geol., vol. 35, p. 607–630. Also a *Reply*, vol. 37 (1929) p. 64–75.

Bryan, Kirk (1940) *The retreat of slopes*, Assoc. Am. Geog., Annals, vol. 30, p. 254–268.

Davis, W. M. (1903) *The Mountain Ranges of the Great Basin*, Harvard Coll., Mus. Comp. Zool., vol. 42, p. 127–175.

——— (1905) *The Wasatch, Canyon, and House ranges*, Harvard Coll., Mus. Comp. Zool., vol. 49, p. 17–56.

Dorf, Erling (1942) *Upper Cretaceous floras of the Rocky Mountain region*, Carnegie Inst. Washington, Pub. no. 508, p. 83–159.

——— (1940) *Relation between floras of type Lance and Fort Union formation*, Bull. Geol. Soc. Am., vol. 51, p. 230.

Eardley, A. J. (1938) *Sediments of Great Salt Lake*, Am. Assoc. Petrol. Geol., Bull., vol. 22, p. 1305–1411.

——— (1939) *Structure of the Wasatch-Great Basin region*, Geol. Soc. Am., Bull., vol. 50, p. 1277–1310.

———, and Haas, Merrill (1936) *Oil and gas possibilities in the Great Salt Lake Basin*, Utah Acad. Sci., vol. 13, p. 61–80.

———, and Hatch, R. A. (1940a) *Pre-Cambrian crystalline rocks of north-central Utah*, Jour. Geol., vol. 48, p. 58–72.

——— (1940b) *Proterozoic (?) rocks in Utah*, Geol. Soc. Am., Bull., vol. 51, p. 795–844.

Gale, H. S., and Richards, R. W. (1909) *Preliminary report on the phosphate deposits in southeastern Idaho and adjacent parts of Wyoming and Utah*, U. S. Geol. Survey Bull. 430, p. 526–529.

George, W. O. (1924) *The relation of the physical properties of natural glasses to their chemical composition*, Jour. Geol., vol. 32, p. 353–372.

Gilbert, G. K. (1890) *Lake Bonneville*, U. S. Geol. Survey, Mon. 1, 438 pages.

——— (1928) *Studies of Basin-Range structure*, U. S. Geol. Survey, Prof. Paper 153, 91 pages.

Gilluly, James (1928) *Basin Range faulting along the Oquirrh Range*, Geol. Soc. Am., Bull., vol. 39, p. 1103–1130.

——— (1932) *Geology and ore deposits of the Stockton and Fairfield quadrangles, Utah*, U. S. Geol. Survey, Prof. Paper 173, 167 pages.

Goodman, Clark, and Evans, Robley D. (1941) *Age measurements by radioactivity*, Geol. Soc. Am., Bull., vol. 52, p. 528, Fig. 10.

Hayden, F. V. (1869) *Preliminary field report* (3d Ann.), U. S. Geol. Survey Colo. N. Mex., 155 pages.

——— (1870) *Sun pictures of the Rocky Mountain scenery, with a description of the geographical and geological features and some account of the resources of the Great West*, New York, 150 pages.

King, Clarence (1876) *Paleozoic subdivisions on the fortieth parallel*, Am. Jour. Sci., 3d ser., vol. 11, p. 475–482.

———, Emmons, S. F., and Hague, Arnold (1877) *Report of the United States Geological Exploration of the 40th Parallel*, vols. 1, 2, 803 pages.

Kober, L. (1928) *Der Bau der Erde*, Berlin, 175 pages.

Lawson, A. C. (1915) *The epigene profiles of the desert*, Univ. Calif. Dept. Geol., Bull., vol. 9, p. 23–48

Leggette, R. M., and Taylor, G. H. (1937) *Geology and groundwater resources of Ogden Valley, Utah*, U. S. Geol. Survey, W. S. Paper 796-D, p. 99–161.

Love, J. D. (1939) *Geology along the southern margin of the Absaroka Range, Wyoming*, Geol. Soc. Am., Special Papers, no. 20, 134 pages.

Mansfield, G. R. (1927) *Geography, geology, and mineral resources of part of southeastern Idaho*, U. S. Geol. Survey, Prof. Paper 152, 446 pages.

Mathews, A. A. L. (1931) *Mesozoic stratigraphy of the central Wasatch Mountains*, Oberlin Coll. Lab., Bull., n. ser., no. 1, p. 1–50.

Neeley, Joseph (1937) *Stratigraphy of the Sundance formation and related Jurassic rocks in Wyoming and their petroleum aspects*, Am. Assoc. Petrol. Geol., Bull., vol. 21, p. 715–770.

Nolan, T. B. (1943) *The Basin and Range Province in Utah, Nevada, and California*, U. S. Geol. Survey, Prof. Paper 197-D, p. 141–195.

Pack, F. J. (1926) *New discoveries relating to the Wasatch fault*, Am. Jour. Sci., 5th ser., vol. 11, p. 399–410.

Peale, A. C. (1879) U. S. Geol. and Geog. Survey Terr., 11th Ann. Rept., p. 509–646.

Pierce, W. G. (1941) *Heart Mountain and South Fork thrusts, Park County, Wyoming*, Am. Assoc. Petrol. Geol., Bull., vol. 25, p. 2021–2045.

Reeside, J. B. (1929) *Triassic-Jurassic "Red beds" of the Rocky Mountain Region: A discussion*, Jour. Geol, vol. 37, p. 47–63.

Richardson, G. B. (1913) *The Paleozoic section in northern Utah*, Am. Jour. Sci., 4th ser., vol. 36, p. 406–413.

——— (1941) *Geology and mineral resources of the Randolph quadrangle, Utah-Wyoming*, U. S. Geol. Survey, Bull. 923, 54 pages.

Schneider, Hyrum (1925) *A discussion of certain geologic features of the Wasatch Mountains*, Jour. Geol., vol. 33, p. 28–48.

Schoff, S. L. (1938) *Geology of the Cedar Hills, Utah*, Ohio State Univ., Abstracts of Doctors' Dissertations, no. 25, p. 375–386.

Spieker, E. M. (Unpublished) *Late Mesozoic and early Cenozoic history of central Utah*, U. S. Geol. Survey.

———, and Reeside, J. B., Jr. (1926) *Upper Cretaceous shoreline in Utah*, Geol. Soc. Am., Bull., vol. 37, p. 429–438.

———, and Schoff, S. L. (1937) *Orogenic chronology of central Utah* (Abstract), Geol. Soc. Am., Pr. 1936, p. 104.

Stanton, T. W. (1893) *The Colorado formation and its invertebrate fauna*, U. S. Geol. Survey, Bull. 106, 288 pages.

Stokes, W. L. (1941) *Stratigraphy of the Morrison formation and related deposits in and adjacent to the Colorado Plateau*, Ph.D. thesis, Princeton Univ.

Veatch, A. C. (1907) *Geography and geology of a portion of southwestern Wyoming*, U. S. Geol. Survey, Prof. Paper 56, 178 pages.

Wegemann, C. H. (1915) *The Coalville coal field*, U. S. Geol. Survey, Bull. 581, p. 161–184.

Wheeler, H. E. (1943) *Lower and Middle Cambrian stratigraphy in the Great Basin area*, Geol. Soc. Am., Bull., vol. 54, p. 1781–1822.

Williams, J. Stewart (1939) *"Park City" beds on southwest flank of Uinta Mountains, Utah*, Am. Assoc. Petrol. Geol., Bull., vol. 23, p. 82–100.

———, and Hanson, Alvin M. (1942) *Phosphate reserves of Utah*, Utah Agric. Ex. Sta., Bull. 304, p. 5–24.

———, and Masey, G. B. (1941) *The Cambrian section in the Logan Quadrangle, Utah and vicinity*, Am. Jour. Sci., vol. 239, p. 276–285.

Wolcott, C. D. (1908) *Nomenclature of some Cambrian Cordilleran formations*, Smithson. Misc. Coll., vol. 53, no. 5, p. 167–230.

Wood, H. E., 2d, *et al.* (1941) *Nomenclature and correlation of the North American continental Tertiary* Geol. Soc. Am. Bull., vol. 52, p. 1–48.

University of Michigan, Ann Arbor, Mich.

Manuscript Received by the Secretary of the Society, November 23, 1942.
Project Grant 330/40.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 55, PP. 895-904, 2 PLS., 1 FIG. JULY 1944

MT. WALDO BATHOLITH AND ASSOCIATED IGNEOUS ROCKS, WALDO COUNTY, MAINE

BY JOSEPH M. TREFETHEN

CONTENTS

	Page
Abstract	895
Introduction	896
Acknowledgment	897
Gneissic granite	897
General description	897
Structure	898
Origin of the foliation	898
Associated pegmatites	898
Diorite	900
Waldo batholith	900
General description	900
Planar flow structure	900
Schlieren	901
Linear structure	901
Fracture pattern	901
Structural interpretation	902
References cited	903

ILLUSTRATIONS

Figure	Page
1. Index map	896

Plate	Facing page
1. Structures in Mt. Waldo batholith	902
2. Structural map of the Mt. Waldo batholith	904

ABSTRACT

The central part of the Bucksport, Maine, topographic map area is underlain by a medium-textured, porphyritic, biotite granite, termed here the Mt. Waldo batholith. The age of the granite has not been determined but is probably either Devonian or Carboniferous. The pluton is roughly circular in plan, with a diameter of approximately 10 miles. Planar flow structures in the granite tending to curve parallel to the margins of the intrusion about the northern end stand steep to vertical. Within the intrusion, the planar flow structure averages about north in strike and vertical in dip. Linear structure is not developed. On the north, northeast, and south, the older metamorphic wall rocks, probably Paleozoic, are crosscut by the granite; elsewhere, the structure of the wall rocks tends to conform to the outlines of the intrusion. Close to the intrusion, the axes of the minor folds in the wall rocks are steeply inclined or vertical; farther from the margins, the pitch of minor folds decreases. The overlying rocks were perhaps pushed bodily upward by the intrusion.

Intrusives older than the granites of the Mt. Waldo batholith are both granitic and more basic. Both groups are of undetermined age. The earlier gneissic granites are foliated and locally show linear structure consisting of biotite streaking on the foliation planes. This type outcrops in small masses. Pegmatite and aplite dikes are present, some older and others younger than the batholith granites. Diabase dikes are the latest intrusions in the area.

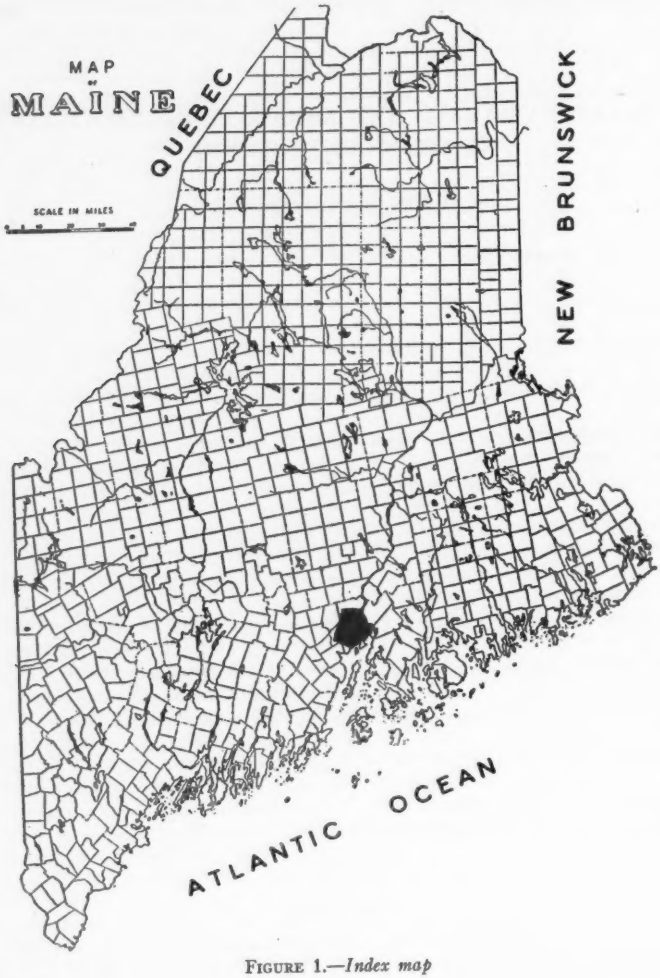


FIGURE 1.—*Index map*

INTRODUCTION

This paper reports partial results of field work carried on in Waldo County, Maine, chiefly in the field season of 1937, supplemented by observations gathered at intervals in subsequent field seasons. The area covered includes part of the Bucksport, Maine, topographic map and parts of adjacent quadrangles to the south and west (Fig. 1).

Little previous geological work has been done in the area. The outline of the chief granite mass, which occupies the central portion of the area, was determined in

reconnaissance by Smith, Bastin, and Brown (1911). The granites have been inspected in the field and in thin section by T. Nelson Dale (1923). Work on the formations which contact the granite massif has been published by Perkins and Smith (1925), although their studies were not carried into the vicinity of the Waldo batholith.

The central portion of the area studied surrounds the highest peak, Mt. Waldo (1062 feet), and is underlain by a medium-grained porphyritic biotite granite, which is intrusive into the surrounding metamorphic rocks. Discussion of the metasediments is withheld pending further work. As is true of many batholiths, several varieties of granite are present. These are thought to be, in this case, at least, closely associated in time as well as space. Preceding and following the intrusion of the Mt. Waldo granites, other igneous rocks were intruded into the region.

ACKNOWLEDGMENT

The writer wishes to acknowledge his indebtedness to the Coe Research Fund of the University of Maine for assistance in carrying out the field work and to Weston S. Evans, Jr., for drafting the map.

GNEISSIC GRANITE

GENERAL DESCRIPTION

The oldest intrusive recognized in the region is the fine- to medium-textured, nonporphyritic, well-foliated biotite gneissic granite. It ranges from dark to medium gray, though at places stained by iron oxides. Locally, the gneissic granite contains inclusions of biotite gneiss and schist. This gneissic granite and its variants are well exposed in the town of Winterport. Few areas of any considerable size of the gneissic granites are exposed in the map area. For the most part, this granite occurs as dikes and sills in the injection gneiss zone which margins the Waldo batholith on the northwest. The injection gneiss into which the gneissic granite is intruded is the migmatite designated as Knox gneiss by Perkins and Smith (1925, p. 225–226) in the area to the southwest. Scattered and discontinuous outcrops of this rock type make exact correlation from place to place impossible. The gneissic granite is cut by the granites of the principal massif.

In describing the intrusions in the Knox gneiss, Perkins and Smith (1925, p. 225–226) describe the rock as

> "biotite granite gneiss which intrudes the sedimentary gneiss in dikes and irregular bodies. Most of these intrusions cut the sediments parallel to the foliation and are themselves foliated in the same direction. Where the intrusion departs from the strike, however, its own foliation remains parallel to the regional structure."

Perhaps the largest and certainly the best exposed mass of the gneissic granite is about a mile north of Winterport Village, where the rock has been quarried for road metal. In the vicinity of Winterport Station there are also good exposures. Relict structures suggesting bedding locally appear in the gneissic granite as, for example, in part of the exposures along the railroad south from Winterport Station. However, most of the rock is clearly transgressive and intrusive.

STRUCTURE

Foliation due to parallel plates of biotite characterizes this type of granite. Lineation is locally present, also, and is shown by streaking of biotites along the foliation surfaces. One of the best places to examine the foliation and lineation is in the abandoned road metal quarry a quarter of a mile west of the main highway, a mile north of the village of Winterport. Here the foliation strikes N. 75°–85° W. and dips 20°–30° N. The lineation pitches 20° in a northeasterly direction. The foliation of the gneissic granite essentially parallels that of the wall rock, and stringers of the granite, tending to become pegmatitic, penetrate the country rock. Aplite dikes are also abundant in some of the exposures of the gneissic granite as, for example, in the road-metal quarry just mentioned. Some of these have disconformable foliation.

ORIGIN OF THE FOLIATION

The origin of the foliation of the gneissic granite is difficult to establish. To the writer it is questionable if well-defined foliation ever develops in a massive granite as a result of post-consolidation deformation. However, it may well be that the foliation shown by the gneissic granite has been accentuated by regional stresses after the consolidation of the magma.

Foliation in some of the aplitic dikes which penetrate the gneissic granite is at an angle to the dike walls and also to the foliation of the gneissic granite. Balk (1937, p. 52–54) has suggested that regional extension has caused such orientation to develop prior to final consolidation of the magma, hence as a primary structure. The writer accepts this as the best explanation of the disconformable structures found in this region. If the gneissic granites were intruded into a country rock, well heated, and at the same time subjected to deforming stresses, the magma might come essentially to rest prior to consolidation and develop oriented texture by slight movements, depending on the tectonic forces affecting the region. Quite possibly the gneissic granite was intruded during the late stages of diastrophism, the earlier emplaced bodies had their foliation determined by regional forces acting during emplacement, and the later associated dikes had their discordant foliation determined by dilation.

ASSOCIATED PEGMATITES

Associated with the gneissic granites throughout the region are pegmatites occurring in pods, lenses, stringers, and dikes. In part, these are cut by the gneissic granite, in part are given off from it, and in part cut the granite. These are simple pegmatites (Landes, 1933) consisting of quartz, microcline, and minor biotite. Their texture is coarse and irregular. They carry none of the rarer pegmatite minerals found in the well-known pegmatites of other parts of the State.

Small pegmatite intrusions are common where there is no granite of this type exposed, intimately interleaved with the schists and gneisses which constitute the country rock. The foliation of the containing rocks is strongly developed in many cases and generally arches around the pegmatite lenses, which are thus concordant

bodies. This pinch and swell structure, well illustrated by Bastin (1911, p. 11–12), is general in Maine areas of intimate intrusion. The lenses at many places are strung along like beads on a string. Seen in favorable exposures, many are elliptical in cross section, as well as in longitudinal section, and generally with a decided rake to the lenses. Bastin (1911, p. 11) states:

> "Another feature highly characteristic of pegmatite masses in foliates is their tendency to swell and thin along their trends so as to form virtually a series of connected lenticules. The contact between pegmatite and foliate is in nearly all areas very sharp, whether the pegmatite lies parallel to or cuts across the folia and whether its mass is large or small. In many places the pegmatite intrusion is so intimate that the bordering schist becomes an injection gneiss. . . . Since the pegmatite frequently cuts across the folia of the sedimentary schists and does not notably change the latter along the contacts; it is plain that the foliation of the schists is not a contact effect of the pegmatite intrusion. It is to be attributed mainly to regional metamorphism previous to the pegmatite intrusions."

It is worth noting, however, that the arching effect of the foliation caused by the bulge of the pegmatite lens dies out a short distance from the lens. Whereas directly at the margin the curvature is marked and notable, a few inches away the folia are undisturbed. There has been thickening of the beds between the lenses and thinning at the bulges. Bastin has assumed that the regional metamorphism and foliation antedate the pegmatites because some of them cut across the folia without disturbing them. For the discordant portions of the pegmatites that is undoubtedly the case. However, for those lenses and pods which are pinched entirely apart or maintain only slender connections, the mechanism of intrusion is difficult to visualize if it is held that they were intruded in such forms. The immediate fading out of the deformation of the surrounding schists about the lenses the with thickening and thinning means rock flowage. Such rock flowage implies that the foliation around the lenses is contemporaneous with the intrusion. If this is true, regional squeezing, metamorphism, and consequent foliation must have accompanied a part of the pegmatite injections. They are in part, then, syntectonic intrusions that have been forced into somewhat mobile media, squeezed into pods and lenses and in some cases into isolated masses prior to final consolidation. Somewhat later invasions have crosscutting relationships. Many quartz stringers and lenses are associated with the pegmatites. In some instances veins of quartz lead out from the pegmatite lenses into the adjacent rock. In the Bucksport map area, just northeast of area shown in Figure 1, these occurrences of pegmatite in the gneisses and schists are well exposed along the high tension line running north from Jacob Buck Pond Mtn. over the shoulder of King's Mtn., and in the adjacent outcrops. The quartz and pegmatite lenses and pods have their long axes rather uniformly inclined to the northeast, angles of pitch averaging about 20°–30°. This is also the approximate strike and pitch of the axes of the minor folds and probably marks the direction of easiest relief and maximum elongation of the region.

In the same area a great many quartz veins and pegmatite stringers show intricate folding, which in part might be interpreted as ptygmatic. However, inasmuch as some of these are apophyses, they are later than the folding, although they are apparently deformed.

It is the author's present opinion that much of the zone mapped as the Knox gneiss

in the Liberty quadrangle by Perkins and Smith, and which extends across the Bucksport quadrangle, is underlain by the gneissic granites, portions of which are locally exposed, and from which dikes and pegmatites have penetrated into the overlying rocks. The intrusion was deep-seated and at a time when the region was under stress, and it continued after the major diastrophism was accomplished.

DIORITE

Well exposed on Gross Point, and locally in the southeastern part of Verona Island, are outcrops of diorite. This diorite is irregular in texture and composition. Hornblende is the principal dark constituent. In places the diorite is schistose. It is cut by many minor granitic and pegmatitic intrusions which are comagmatic in part, at least, with the granite intrusions east of Penobscot Bay mapped by Smith, Bastin, and Brown (1907). The diorite is interpreted as an early phase of the granite intrusions. The granites adjacent to Penobscot Bay on the east, including the large Lucerne batholith, not yet mapped, are largely similar in type and relationships to the Mt. Waldo batholith and probably belong to the same magma series.

WALDO BATHOLITH

GENERAL DESCRIPTION

The principal intrusion of the region is the Waldo batholith. This granite underlying the central portion of the Bucksport topographic map area, and holding up the highest hills, is a light-gray to medium-gray porphyritic biotite granite of medium to coarse texture, with potash feldspar phenocrysts up to $1\frac{1}{2}$ inches long. Locally hornblende accompanies the biotite. Thin dikes of pegmatite and aplite are found in most exposures, and knots are sparsely present in many exposures. Schlieren are not abundant, although locally developed.

PLANAR FLOW STRUCTURE

In the majority of the outcrops a planar flow structure is present. It consists chiefly of subparallel arrangement of the feldspar phenocrysts and, less distinctly, of the biotite. In some outcrops this planar structure is very well marked; in others it is only faintly developed, or absent. The strength of the foliation, however, is not related in any obvious way to its proximity to the margin of the intrusion. Throughout the intrusion, the planar structure is vertical or very steep (Pl. 2). The orientation of the foliation is influenced by the margins of the intrusive in varying degrees. In the south and southwest parts of the area it is only locally parallel to the contact. In the north and east portions of the massif it is distinctly parallel to the contact. Toward the margins in the northeast portion, it swings around from north-northeast to west by north along the north margin, parallel to the contact. Along this northern contact the granite crosscuts the structure of the country rock. The actual contact, however, is not exposed. This westward deflection of the foliation along the northern contact is marked within the massif for a distance of from 1 to 2 miles from the margin.

The foliation in the quarry on the north spur of Mt. Waldo averages nearly north-south. On the crest of this spur, northwesterly strikes prevail, which swing more strongly west to the north. On Mosquito and Heagan mountains the foliation also strikes northerly, varying but few degrees in either direction. On Treat Hill, just east of Frankfort Village, the foliation strikes westerly, averaging about west-northwest, and on the hills on both north and south sides of the road leading west from Frankfort the westerly trend of the foliation is marked. The foliation is very steep, averaging about vertical.

Throughout most of the interior of the granite area the foliation is marked, averages nearly north in strike, and is vertical or very steep.

SCHLIEREN

Schlieren, showing a platy arrangement, and in general parallel with the foliation, are locally observable. They cannot be traced more than a few feet in most exposures. Lithologically they appear to be identical with the granites containing them, except for the greater concentration of biotite and hornblende. In the abandoned quarry on Mosquito Mountain, Frankfort, they are particularly well developed and show some distinctive peculiarities. On both the east and west sides of the quarry, dikes only slightly younger than the main mass of granite are present, very strongly streaked with schlieren, parallel to the margins of these dikes. Other schlieren areas in this quarry appear to have dikelike relations, in that the foliation of the schlieren crosscuts the foliation of the granite in section, although paralleling it in general in plan. These schlieren dikes appear in no way different from normal schlieren, except for their crosscutting relationships. They are cut by thin aplite dikes which strike approximately normal to the foliation of the granite. On the west side of the quarry, one such schliere appears to have been faulted, but there is no trace of any fracture surface left; it is perfectly sealed and healed (Pl. 1, fig. 1).

Along the strike of one of these schlieren-rich zones, the dikelike aspect becomes more and more apparent, and the schlieren zone gradually becomes a normal aplite dike. There appears to be no gradation from knots to schlieren in this quarry. The knots are abundant; probably over 1000 could be counted within the quarry limits, and although various stages of reaction with the granite appear to be distinguishable none of them is drawn out into anything approaching a schliere. The knots are of several types and are probably inclusions.

LINEAR STRUCTURE

Linear structure in the granite is absent over most of the area. At the Waldo quarry, along the margins of a central finer-grained dike some 200 feet wide that runs along the axis of the quarry, are numerous "knots". Some of these are elongated and pitch about 20°N. Elsewhere, as on Heagan Mountain, knots are strung out lengthwise in plan, but measurements of pitch are impossible because of a lack of vertical exposures.

FRACTURE PATTERN

Throughout most of the area of granite, steep joints are not abundant. At many outcrops, however, measurements have been made on the more prominent joints. These are dominantly steeply inclined to vertical and are strongly westerly in strike

throughout most of the area. The same type of joint with similar attitude prevails throughout the adjacent sedimentary and metamorphic wall rocks for distances of at least several miles from the Waldo batholith. Around the north margin of the batholith, where the deflection of the foliation is so strongly marked, there is no regular joint fan, as has been described for many granitic intrusions; but here, as elsewhere, the majority of the joints strike west-northwest.

While joint fillings, in the form of narrow pegmatite and aplite dikes, trend in various directions, the majority of these are in the westerly striking joint set (Pl. 1, fig. 2). Locally movements of small magnitude are apparent along some of the steep joints, evidenced by slickensiding and minor offsets of small dikes transecting the joint.

Sheeting, however, is prominent on most of the granite hills. A particularly well developed sheeting is to be seen on the east side of Mosquito Mountain. In general, the sheeting dips a little less steeply than the present erosion slopes and dips quaquaversally away from the hill tops. In the Mt. Waldo quarry, there are several fracture zones associated with the sheeting (Pl. 1, fig. 3). These subhorizontal zones have diagonal cracks closely spaced and probably represent a type of fracture cleavage.

STRUCTURAL INTERPRETATION

The flow structure of the granite indicates that the magma was actively intruded and flowed in a general north-south direction just previous to congelation. On the northeast and north margins, the granite cuts strongly across the structure of the adjacent rock, which must, therefore, have been deformed prior to the invasion of the granite. The contact zone of the batholith appears to be one of weakness, for it is quite generally occupied by valleys. The preference of streams for the contact zones of granite intrusives is common in Maine.

Along the southeast margin, the strike of the wall rocks changes rather abruptly from northeasterly to northwesterly, suggesting drag probably due to the intrusion of the granite. The flow structure in the granite is not strongly developed along the southern margin and is somewhat disconformable. Inasmuch as linear structures are lacking, inferences as to the actual directions of the movements in the planes of foliation are unwarranted. Nor is the attitude of the steep joints and associated fillings to be greatly relied upon in deducing the actual direction of the magma source, because these joints are a regional feature, similar in trend and attitude to those well outside the massif.

In considering plutons with steeply dipping flow structure Balk (1937, p. 87) suggests that (1) The flow structure closed overhead in the form of a dome, or (2) the granite penetrated its roof rock *lit-par-lit* fashion. The evidence here for either case is eroded away. The sheeting is evidently late, because on the southeast slope of Mosquito Mountain the sheeting passes through a vertical diabase dike which has chilled margins. Moreover, under either of the two suggested forms of the intrusion, the sheeting can scarcely be attributed to cooling and shrinkage. Under the first assumption considerable erosion has taken place, and the present exposed surfaces must be well below the original contact zone. Under the second assumption no regular system of sheeting joints paralleling in a cooling surface can be postulated

FIGURE 1. SCHLIEREN, MOSQUITO MTN. QUARRY

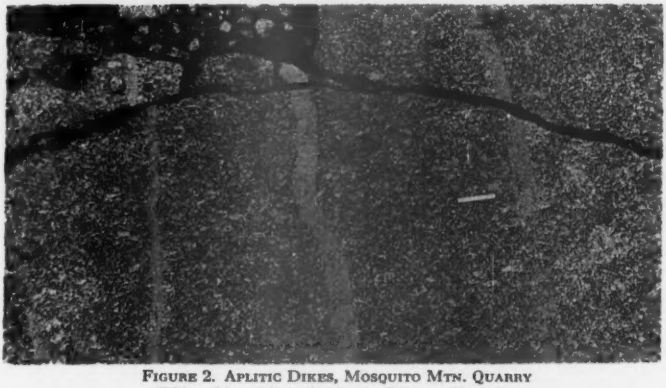


FIGURE 2. APLITIC DIKES, MOSQUITO MTN. QUARRY
White object 2¾ inches long. Dikes strike west.

FIGURE 3. FRACTURE ZONES ASSOCIATED WITH SHEETING
Mt. Waldo Quarry.

STRUCTURES IN MT. WALDO BATHOLITH

The fracture cleavage zones in the sheeting on Mt. Waldo suggest expansive movements. It is suggested, therefore, that the exfoliation joints are the result of expansion due to unloading through erosion.

The steepening of the axes of minor folds adjacent to the granite suggest that stoping and/or assimilation are not the sole answers to the emplacement of this granite body. It is equally possible, although difficult to demonstrate, that it has upthrust the cover in some considerable measure. Before consolidation was finally complete, the magma found easier relief to the south, and the general north-south trend of the foliation was impressed. As the magma crystallized into a potentially dilatent state—that is when the crystals became so closely packed that deformation was possible only with an increase in volume—the mass was subject to cracking, and the cracks being zones of low fluid pressure drew fillings into themselves. Such fillings are the aplitic and pegmatitic dikes, which are normal to the elongation.

The basic dikes, found at scattered localities, are the youngest igneous rocks of the region.

REFERENCES CITED

Balk, Robert (1937) *Structural behavior of igneous rocks*, Geol. Soc. Am., Mem. 5, 177 p.

Bastin, E. S. (1911) *Geology of the pegmatites and associated rocks of Maine*, U. S. Geol. Survey, Bull. 445, Pl. I.

Dale, T. Nelson (1923) *Commercial granites of New England*, U. S. Geol. Survey Bull. 738, p. 257–261.

Landes, K. K. (1933) *Origin and classification of pegmatites*, Am. Mineral., vol. 18, p. 33–56, 95–103.

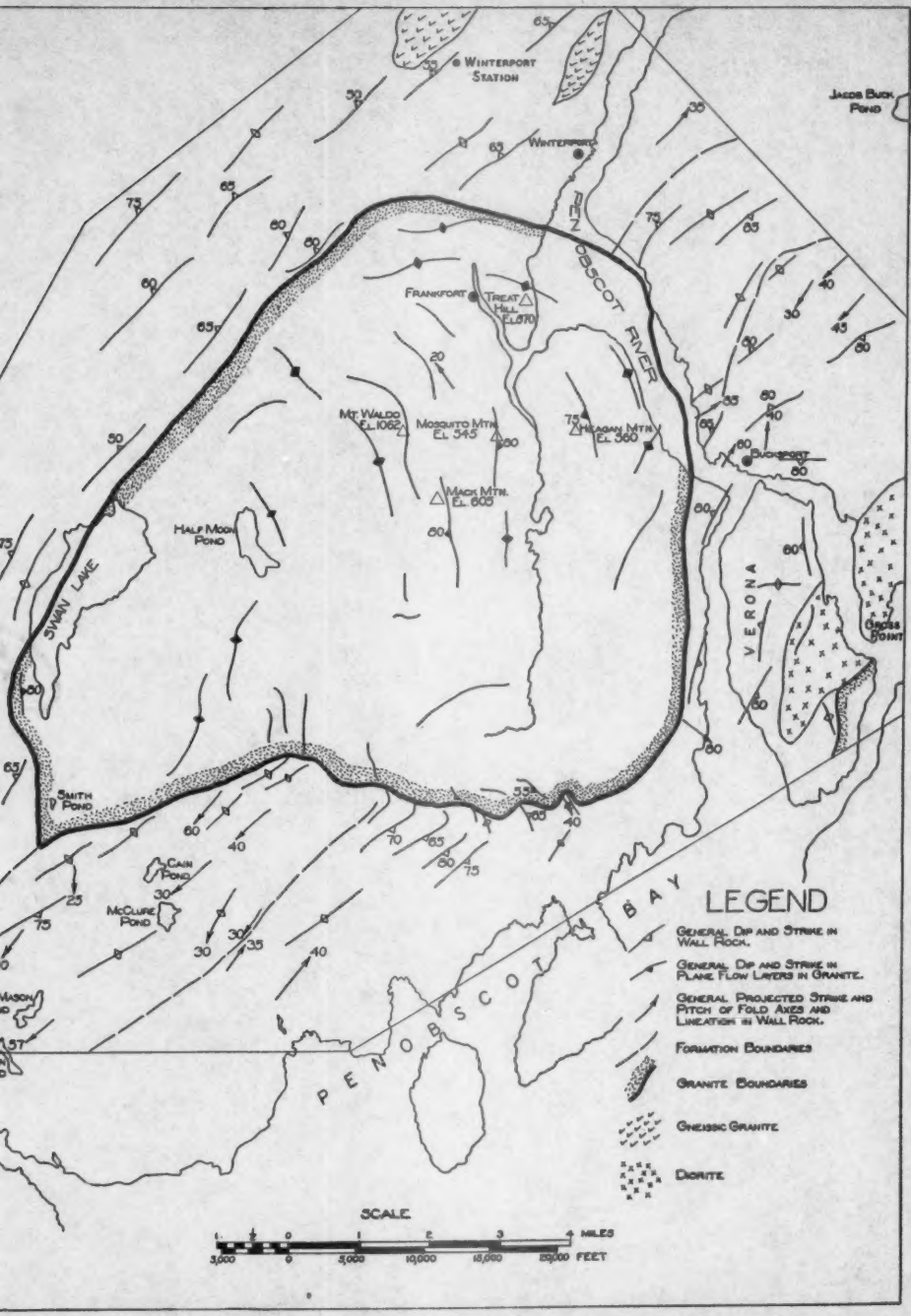
Perkins, E. H., and Smith, E. S. C. (1925) *A geological section from the Kennebec River to Penobscot Bay*, Am. Jour. Sci., 5th ser., vol. 9, p. 204–228.

Smith, G. O., Bastin, E. S., and Brown, C. W. (1907) *Description of the Penobscot Bay quadrangle*, U. S. Geol. Survey, Geol. Atlas, Folio, no. 149.

University of Maine, Orono, Maine
Manuscript Received by the Secretary of the Society, February 5, 1943

SWAN LAKE
SMITH'S POND
NICHOLS POND
HURDS POND
UPPER MASON POND
MASON POND
25
73
60
65
63
70
22
75
20
57

ST



STRUCTURAL MAP OF THE MT. WALDO BATHOLITH

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 55, PP. 905-920, 1 PL., 1 FIG. JULY 1944

PETROLOGY OF THE PAWTUCKAWAY MOUNTAINS, NEW HAMPSHIRE

BY CHALMER J. ROY AND JACOB FREEDMAN

CONTENTS

	Page
Abstract	906
Introduction and acknowledgments	906
Previous geological work	906
General geology	908
New Hampshire magma series, Winnipesaukee gneiss (?)	908
White Mountain magma series, Pawtuckaway complex	908
Porphyritic gabbro	908
Diorites	909
Microscopic petrography of diorites	910
Augite diorite	910
Hornblende diorite	910
Foliated hornblende diorite	911
Hornblende-biotite diorite	911
Porphyritic diorite	911
Monzonites	911
Microscopic petrography of monzonites	912
Coarse-grained monzonite	912
Fine-grained monzonite	912
Dike rocks	912
Microscopic petrography of dike rocks	912
Monzonites	912
Lamprophyres	912
Aplites	912
Structure	913
Winnipesaukee gneiss (?)	913
Pawtuckaway complex	913
Diorite body	913
Outer monzonite arc	913
Middle Mountain complex	914
Dikes	914
Joints	914
Age relations	915
Mode of intrusion	915
General statement	915
Diorite body	915
Forceful injection	915
Piecemeal stoping	916
Cauldron subsidence	916
Monzonite bodies	916
Forceful injection	916

Piecemeal stoping 916
Cauldron subsidence 916
Contrasts with other areas 917
Sequence of events 917
Relation of intrusion to differentiation 918
Conclusions 918
References cited 918

ILLUSTRATIONS

Figure — Page
1. Index map 907

Plate — Facing Page
1. Geologic map of the Pawtuckaway Mountains 920

ABSTRACT

The Pawtuckaway Mountains are located in Deerfield and Nottingham townships, Rockingham County, in southeastern New Hampshire.

The rocks of the area constitute an intrusive complex belonging to the White Mountain magma series which was injected into the Winnipesaukee gneiss (?) of the New Hampshire magma series. The complex consists of a variety of diorites and monzonites with minor amounts of gabbro. Aplitic, monzonitic, and lamprophyric dikes cut all the major rock types. Some aplite dikes are younger than lamprophyre dikes which they intersect, but the age relationships of the monzonite dikes are unknown.

The oldest units of the complex are the gabbros followed in order by the diorites, monzonites, and dike rocks. Diorites and monzonites constitute the major units of the complex and are believed to have formed from one parent magma differentiating at depth, although the variations in the diorite body may have resulted from differentiation in place.

Cauldron subsidence or ring-dike injection is favored as the method of emplacement on the basis of the circular plan of the complex, the arcuate pattern of the monzonite bodies, and the general structural and petrographic similarity to other ring-dike areas of the White Mountain magma series.

INTRODUCTION AND ACKNOWLEDGMENTS

The Pawtuckaway Mountains are located in the towns of Deerfield and Nottingham, Rockingham County, New Hampshire (Fig. 1). The mountain area is elliptical in plan, being approximately 2 miles long north and south and 1½ miles wide east and west. The Pawtuckaway State Park includes most of the mountain area. The relief ranges from 300 to 500 feet.

This study was begun by Roy, who spent parts of the summer of 1935 and the following winter mapping the geology. With the hand specimens, thin sections, field notes, and maps prepared by Roy, Freedman spent the academic year 1939–1940 on a laboratory study of the rocks and some time in the field to complete the mapping. Acknowledgments are gratefully accorded Professors Larsen and Billings of Harvard University for helpful suggestions. Very helpful aid was also given by J. B. Lyons in the laboratory and by Katharine Fowler-Billings and T. Botinelly in the field. Chemical analyses were made by F. A. Gonyer.

PREVIOUS GEOLOGICAL WORK

This study is the first published report on the geology of the area. C. T. Jackson (1844, p. 50) mentioned the Pawtuckaway Mountains as being composed of "sienite"

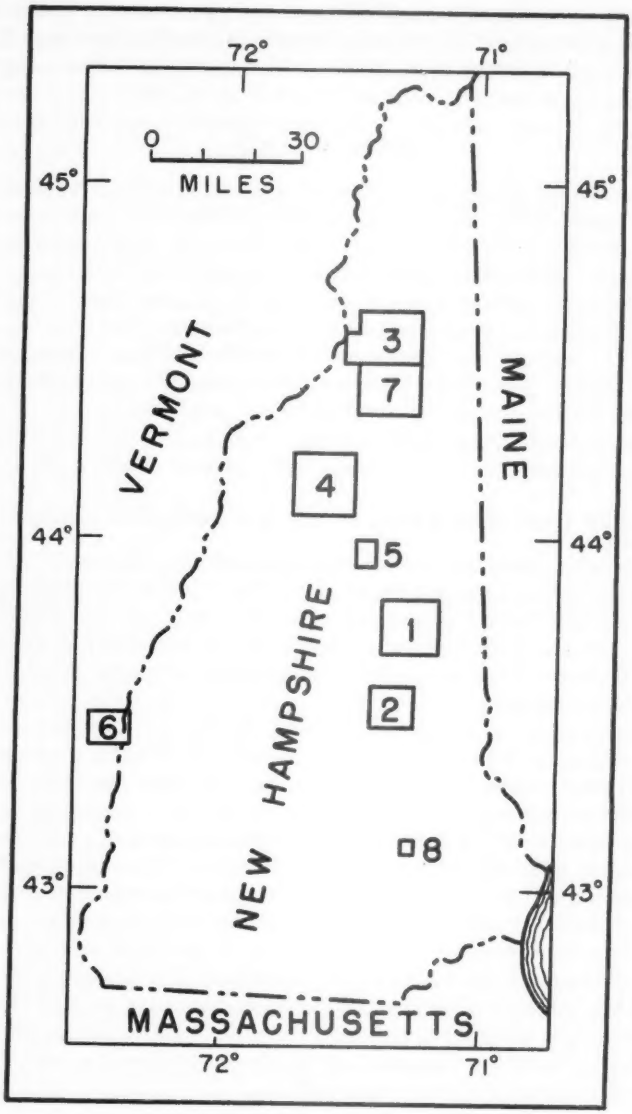


FIGURE 1.—*Index map*

Showing known areas of rocks belonging to the White Mountain magma series.

1. Ossipee Mountains
2. Belknap Mountains
3. Percy quadrangle
4. Franconia region
5. Mount Tripyramid
6. Ascutney Mountain
7. Pliny region
8. Pawtuckaway Mountains

and discussed the great depth of weathering in a quarry northeast of Middle Mountain. C. H. Hitchcock (1877; 1878) also referred to the rocks as syenites. E. S. C. Smith (1922, p. 128) termed the rocks "an alkaline syenite intrusive into a granitic gneiss."

GENERAL GEOLOGY

The Pawtuckaway Mountains are composed of plutonic, alkaline rocks of the White Mountain magma series, probably Mississippian, intrusive into the Winnipesaukee gneiss (?) (Hitchcock, 1877, p. 553) of the New Hampshire magma series, probably late Devonian. Although the gneiss in this area is unlike the Winnipesaukee gneiss at the type locality (Billings, personal communication), it was considered to be the same by Hitchcock, and the present authors retain the name. The alkaline complex is circular like areas of similar igneous rocks of the White Mountain magma series in the New England petrographic province. The dominant rocks are diorites and monzonites, but the complex contains rocks which range from gabbro to syenite. The monzonites underlie the higher portions of the area; the valleys are underlain by diorite. Lamprophyre, monzonite, and aplite dikes are abundant (Pl. 1).

NEW HAMPSHIRE MAGMA SERIES, WINNIPESAUKEE GNEISS

In this area the gneiss is gray or white, holocrystalline, equigranular, and is composed mainly of quartz, potash feldspar, and biotite. The degree of foliation is distinctly variable. The rock ranges from apparently nonfoliated, granitoid to highly schistose. In general the degree of foliation is directly related to the abundance of biotite. Pegmatite dikes, mainly small and composed of quartz, potash feldspar, and biotite, are common.

Microscopic study of a moderately foliated (biotite-poor) specimen shows the mineralogical composition to be quartz 30 per cent; potash feldspar, mainly orthoclase and perthite, 60 per cent; oligoclase (An_{17} to An_{12}) about 5 per cent; biotite and minor amounts of sericite and magnetite about 5 per cent. The quartz grains are clear and anhedral; a few are poikilitic in perthite suggesting contemporaneous crystallization. The bulk of the quartz is interstitial. Some quartz grains show flamboyant extinction. Orthoclase and perthite occur in subhedral to anhedral grains and show slight alteration to sericite. Biotite occurs in irregular grains and shows strong pleochroism, X = greenish-brown, Y and Z = dark brown. The biotite contains small grains of magnetite, probably an alteration product.

The gneiss is a result of the injection of granitic material of the New Hampshire magma series into schistose phases of older rocks (Katz, 1917). In the region of the Pawtuckaway Mountains the granitic material dominates to such an extent that the foliation of the gneiss is distinct only locally.

WHITE MOUNTAIN MAGMA SERIES, PAWTUCKAWAY COMPLEX

PORPHYRITIC GABBRO

The oldest rock in the Pawtuckaway complex is a porphyritic gabbro found in three small areas. It is black where weathered but resinous brown with black py-

roxene phenocrysts where fresh. The groundmass is mainly medium-grained; the phenocrysts are distinct and range up to more than half an inch in diameter. The phenocrysts stand out as knots on the weathered surface.

Microscopic study shows the rock to be holocrystalline, coarsely porphyritic, with the constituents mainly anhedral to subhedral. An approximate mode of the porphyritic gabbro is given in Table 1.

TABLE 1.—*Approximate modes of typical Pawtuckaway rocks*

Type	1	2	3	4	5	6	7	8	9	10
Plagioclase										
Oligoclase	—	74	—	55	47	40	40	5	—	—
Andesine	—		45	—	—	—	—	—	58	—
Labradorite	30	—	—	—	—	—	—	—	—	—
Potash feldspar	—	—	tr	5	10	30	35	85	tr	80
Hornblende	5	3	25	12	18	10	7	3	20	5
Pyroxene	50	20	5	5	5	5	4	1	15	5
Olivine	5	—	4	—	—	—	—	—	—	—
Biotite	7	2	12	18	17	10	9	4	5	—
Quartz	—	—	tr	tr	tr	1	2	1	—	5
Accessories	3	1	9	5	3	4	3	1	2	5
Total	100	100	100	100	100	100	100	100	100	100

1. Porphyritic gabbro.
2. Augite diorite.
3. Hornblende diorite.
4. Hornblende-biotite diorite.
5. Porphyritic diorite.
6. Coarse-grained monzonite.
7. Fine-grained monzonite.
8. Syenite.
9. Lamprophyre.
10. Syenite aplite.

The pyroxene (diopside-hedenbergite, containing approximately equal percentages of the two molecules) is twinned, strongly schillerized, and clouded by grains of secondary magnetite. It is altered in part to brown, basaltic hornblende similar to that to be described under hornblende diorite. Feldspar laths of labradorite (An_{58} to An_{63}) are twinned according to the albite and Carlsbad laws and have normal zoning, of which even the most sodic band is labradorite. The laths are interstitial to the augite grains and embay them. The olivine (forsterite 80 per cent) occurs as irregular grains which are altered to iddingsite, serpentine, and magnetite.

DIORITES

Diorites constitute a major portion of the Pawtuckaway complex, and a number of distinct varieties have been differentiated on the basis of texture, structure, and mineral composition: (1) augite diorite, (2) hornblende diorite, (3) foliated hornblende diorite including diabase, (4) hornblende-biotite diorite, and (5) porphyritic diorite. Each of these constitutes a mappable unit. Microscopic examination confirms the field distinctions in each instance.

The augite diorite is similar megascopically to the porphyritic gabbro except that it is nonporphyritic. The two rocks occur in close association and are therefore not differentiated on the map. The hornblende diorite is black to resinous gray and very coarse-grained. Maximum grain size exceeds 10 mm. The foliated hornblende diorite is similar in color and texture to the hornblende diorite but is distinctive

because of the foliation. Both feldspar and hornblende grains show parallel orientation and segregation into feldspar-rich and hornblende-rich layers producing a marked gneissoid foliation. A related rock is a diabase in which only the feldspar is oriented and segregation is absent. The two rocks are not differentiated on the map because

TABLE 2.—*Chemical analyses of rocks*

SiO_2	57.44	56.12	48.08	50.48
TiO_2	1.49	1.10	2.40	1.45
Al_2O_3	18.80	16.96	18.99	15.34
Fe_2O_3	.63	2.93	2.54	3.84
FeO	6.31	4.01	7.88	7.78
MnO	.08	.16	.06	.20
CaO	3.78	6.50	9.54	8.94
MgO	1.03	3.27	3.39	5.79
Na_2O	5.58	3.67	4.32	3.07
K_2O	3.42	3.76	.61	.97
H_2O-	1.10	1.05	.08	1.89
H_2O	.39		.42	
P_2O_5	.38	.47	1.16	.25
SO_3	.38	—	.64	—
BaO	.04	—	.02	—
F	—	—	—	—
Total	100.85	100.00	100.13	100.00
Less O for F				

1. Coarse-grained monzonite from Boulder Road just east of Round Pond. F. A. Gonyer, analyst.
2. Average of 27 analyses of monzonite from Daly (1933, p. 13).
3. Foliated hornblende diorite from field east of Goodrich Farm. F. A. Gonyer, analyst.
4. Average of 90 analyses of diabase from Daly (1933, p. 18).

they are petrographically the same. The main body mapped as foliated diorite in the eastern part of the area is mainly the gneissoid variety, whereas the smaller area along the road in the western part of the area is diabase. The hornblende-biotite diorite is medium-grained and contains more biotite and less hornblende than the others. The porphyritic diorite is almost pegmatitic in texture, having a white feldspar groundmass and extremely coarse hornblende phenocrysts, some of which attain a length of more than 150 mm.

MICROSCOPIC PETROGRAPHY OF DIORITES

Augite diorite.—The augite diorite is holocrystalline, inequigranular, and coarse-grained. Subhedral, dark, clouded feldspar laths and augite are the essential minerals. An approximate mode is given in Table 1. The feldspars are long laths of normally zoned plagioclase which range from oligoclase to andesine (An_{20}–An_{40}). Coarse and fine albite twinning and Carlsbad twinning are common. The augite is similar to that in the gabbro but is less abundant and does not occur as distinct phenocrysts. It is marginally altered to brown and green hornblende. Scattered grains of biotite are partially altered to magnetite. Apatite is accessory, and occurs in large, scattered crystals, a typical occurrence in all the Pawtuckaway rocks. Magnetite is partly primary and partly a result of alteration of augite, hornblende, and biotite.

Hornblende diorite.—The hornblende diorite is very coarse-grained, holocrystalline, and equigranular. The constituent grains range from subhedral to euhedral. The most conspicuous are hornblende which range up to 10 mm. by 13 mm. An approximate mode is given in Table 1. The feld-

spar is mainly andesine (An_{32} to An_{38}), but some normal and some combinations of normal and reverse zoning, which range from labradorite to calcic oligoclase, are present. The grains are large, and coarse albite twinning and Carlsbad twinning are common. The amphibole is mainly brown, basaltic hornblende although a little green hornblende is present. The optical properties of the brown hornblende are: $\alpha = 1.681$, $\beta = 1.693$, $\gamma = 1.703$, 2V = high, dispersion $r < v$, $Z \wedge c\ 9°$, pleochroism $Z > Y > X$, Z = dark brown, Y = brown, X = light brown. These optical data are not like those of most of the amphiboles of the White Mountain magma series so far published. Smith (1940) described a basaltic hornblende from a gabbro with the following optics: $\alpha = 1.672$, $\beta = 1.682$, $\gamma = 1.689$, 2V = large. As the degree of oxidation of basaltic hornblende varies directly with the nearness of Z to c = 0° (Barnes, 1930, p. 393), oxidation here is not complete. Apatite, magnetite, and feldspar are poikilitic in hornblende and biotite, and hornblende is altered in part to biotite. Hypersthene is the only pyroxene observed. It is colorless in thin section and is altered in part to green hornblende and magnetite. Large grains of brown biotite are scattered through the rock, and smaller grains occur as alteration of the hornblende. The biotite is altered locally to pale-green chlorite. A few irregular grains of olivine with brown iddingsite are scattered through the rock. The optical properties of the olivine are: $\alpha = 1.661$, $\beta = 1.679$, γ 1.704, 2V = 85°, dispersion $r < v$, X = b, Z = a. This olivine has over 80 per cent of the forsterite molecule; that from the White Mountain magma series of other areas contains more of the fayalite molecule (Chapman and Williams, 1935). Hyalosiderite ($Fa_{40}Fo_{60}$) occurs in gabbros, and fayalite ($Fa_{90}Fo_{10}$) is found in the syenites and granites in other bodies of the White Mountain magma series. For comparison, optical properties of hyalosiderite from a gabbro from Mt. Tripyramid, New Hampshire, are presented: $\alpha = 1.721$, $\beta = 1.750$, $\gamma = 1.765$, 2V = 85° ± 2°, $r > v$, weak. The accessories, magnetite and apatite, are more abundant in the hornblende diorite than in the augite diorite and in the gabbro.

Foliated hornblende diorite.—The foliated hornblende diorite is similar in nearly all respects to the normal diorite described above. The principal differences are the orientation of the grains, the general absence of euhedral grains, and the dominance of a feldspar richer in the anorthite molecule. The feldspar is predominantly zoned, ranging from labradorite to andesine.

Hornblende-biotite diorite.—The hornblende-biotite diorite, although holocrystalline, is distinctly finer-grained than most of the diorites. The feldspar is mainly oligoclase (An_{23}), though some grains show combinations of normal and reversed zoning, ranging from andesine to oligoclase. There are also a few grains of orthoclase and microcline. Except for their relative abundance, the hornblende and biotite in this rock are the same as those of the normal diorite. The absence of pyroxene in some specimens of the hornblende-biotite diorite indicates that pyroxene crystals have been completely reworked to hornblende.

Porphyritic diorite.—The porphyritic diorite is a distinctive rock consisting of a white, granular groundmass of feldspar with large elongated phenocrysts of hornblende. The phenocrysts range up to 150 mm. in length and have an average length between 25 mm. and 50 mm. Locally the rock appears to be almost pegmatitic.

An approximate mode of the porphyritic diorite is given in Table 1. The large hornblende phenocrysts have the following optical properties: $\alpha = 1.685$, $\beta = 1.698$, $\gamma = 1.710$, 2V = high, dispersion $r < v$ strong, $Z \wedge c\ 4°$, pleochroism Z = dark greenish-brown, Y = greenish-brown, X = pale green. An amphibole with these properties is a basaltic hornblende with indices slightly higher, birefringence greater, and extinction angle smaller than those of the basaltic hornblende in the hornblende diorite. The feldspar is oligoclase (An_{23} to An_{28}), with a little orthoclase.

MONZONITES

The monzonites of the Pawtuckaway complex are separated texturally into coarse-grained and fine-grained varieties which are distinct enough to constitute mappable units.

The coarse-grained monzonite is light to dark resinous gray and equigranular. The amounts of biotite and hornblende vary, the former usually appearing to be more abundant. A syenitic facies present locally contains predominant potash feldspar and less biotite and hornblende.

The fine-grained monzonite is mainly dark gray or blue gray, equigranular, and similar in mineral composition to the coarse-grained monzonite. Locally it is porphyritic, with phenocrysts of biotite and hornblende. Inclusions of gabbro occur in the fine-grained monzonite at some places.

MICROSCOPIC PETROGRAPHY OF MONZONITES

Coarse-grained monzonite.—In the coarse-grained monzonite, the feldspar is orthoclase, microcline, and slightly sericitized plagioclase with reverse zoning ranging from oligoclase to andesine (An_{42}). The hornblende is green with the following optical properties: $\alpha = 1.672, \beta = 1.690, \gamma = 1.706, 2V = 70°$, dispersion $r < v$, pleochroism Z = greenish-black to dark-olive green, Y = olive green, X = yellowish green, $Z \wedge c$ 11°. The optical properties of the brown biotite, which is similar to that in the diorites and the fine-grained monzonite, are: $\alpha = 1.619, \beta = 1.673, \gamma = 1.673, 2V = 5°$, dispersion $r < v$ strong, pleochroism strong, X = light reddish-brown, Y and Z = dark brown. A biotite with these optical properties is essentially siderophyllite, rich in ferric iron. It differs from a biotite in the granite of the Percy quadrangle (R. W. Chapman, 1935) in having a slightly higher alpha index and strong dispersion of the optic axes. The pyroxene is augite and a few grains of hypersthene. A chemical analysis of the coarse-grained monzonite is given in Table 2.

Fine-grained monzonite.—An approximate mode of the fine-grained monzonite is given in Table 1. The feldspar is potash feldspar including microperthite, and oligoclase (An_{28}), the latter being twinned according to the albite and Carlsbad laws, and slightly zoned and sericitized. The optical properties of the hornblende are: $\alpha = 1.685, \beta = 1.693, \gamma = 1.708$, 2V = high, dispersion $r < v$, $Z \wedge c$ 17°, pleochroism Z = dark greenish-brown, Y = olive green, X = yellowish green. This differs from the hornblende of the coarse-grained monzonite in having slightly higher indices, a larger extinction angle, and a larger axial angle. The accessories are small needles of apatite, round grains of magnetite, rare grains of zircon, and interstitial quartz.

DIKE ROCKS

Monzonite, lamprophyre, and aplite dikes, mainly small and discontinuous, are common throughout the area. The monzonite dikes are similar to the fine-grained monzonite described above. The lamprophres are black, coarse- to fine-grained, equigranular or porphyritic rocks, essentially diorites. The phenocrysts in the porphyritic lamprophyres are augite and biotite. The aplites are white or blue gray with equigranular, saccharoidal, or dense texture. Rare, medium- to coarse-grained syenitic and monzonitic rocks are mapped as aplites.

MICROSCOPIC PETROGRAPHY OF DIKE ROCKS

Monzonites.—In the monzonite dikes, the plagioclase is zoned oligoclase to andesine (An_{27} to An_{38}), and the potash feldspars are orthoclase, microcline, perthite, and antiperthite. The feldspars are slightly altered to sericite, and the cores of some of the zoned grains are altered to calcite. The pyroxenes, augite, and hypersthene are altered to hornblende, magnetite, and bastite. Brown biotite constitutes up to 10 per cent of some monzonite dikes. Magnetite, apatite, and zircon are accessory.

Lamprophyres.—The feldspar, andesine (An_{38}) and a few grains of microcline, occurs in unoriented laths, which in the fine-grained dikes are small to microlitic. The hornblende is green and greenish brown. It is in part an alteration product of augite but is mainly primary. Some of the hornblende is altered to magnetite along cleavage planes, and a little has been schillerized. The pyroxene is augite. Small irregular patches of green to brown biotite are slightly altered to magnetite. A little accessory magnetite and apatite are poikilitic in biotite.

Aplites.—An approximate mode of the aplites is given in Table 1. The feldspars are orthoclase,

microcline, microperthite, antiperthite, and a little oligoclase (An_{18}). The small amount of amphibole, 5 per cent or less, is hornblende and, in one dike, hastingsite; the pyroxenes are augite and hypersthene. Some of the quartz occurs as clear, interstitial grains, and some is present in myrmekitic intergrowths. A little biotite is present in small scattered grains, and the accessories are apatite, magnetite, and zircon.

STRUCTURE

WINNIPESAUKEE GNEISS (?)

The gneiss strikes northeast-southwest, parallel to the regional structural trend in southeastern New Hampshire (Katz, 1917). The area immediately surrounding the Pawtuckaway complex is densely wooded, low, and in part swampy. Careful search was made for gneiss exposures near the intrusive complex, and the attitude of the schistosity, where obtainable, is mapped. The foliation of the gneiss does not conform to the outline of the intrusive complex.

PAWTUCKAWAY COMPLEX

Diorite body.—Although several mappable varieties of diorite are recognized, they constitute a single intrusive unit. The diorite is the oldest portion of the complex, and it must have been intruded into the older gneiss as an irregularly elliptical body. The outer arc of monzonite was later injected between the diorite and the gneiss for three-quarters of its circumference, leaving the diorite in contact with the gneiss only in the western part of the area (Pl. 1).

The contact between the diorite and gneiss is exposed at only one place—just north of the road which enters the area from the west. The contact strikes N. 5°E. and dips 55° N., indicating that it dips outward. The strike of the foliation in the diorite near this point is parallel to the contact, but the dip is unobtainable; however, along the road to the southeast the foliation dips steeply toward the center of the complex.

Perhaps the most significant structural feature of the diorite is the foliation which is well developed in certain portions. The major area of foliated diorite lies between the outer monzonite arc and the Middle Mountain unit. Except for the foliation the rock here is essentially the same as the normal hornblende diorite.

The strike of the foliation throughout most of this area is essentially parallel to the outer and inner contacts with the monzonite. The dips, where obtainable, are consistently toward the center of the complex and range from 40° to 80°.

Outer monzonite arc.—The main body of coarse-grained monzonite is intrusive between the diorite body and the Winnipesaukee gneiss (?) (Pl. 1). Cropping out in an arcuate pattern, the coarse-grained monzonite surrounds the diorite body for three-quarters of its circumference in a manner similar to ring-dikes described in other New Hampshire areas (Kingsley, 1931; Modell, 1936; Williams and Billings, 1937; Chapman, 1935; and others). The approximate contact between the coarse-grained monzonite and the adjacent rocks can be determined in numerous places within 10 or 20 feet, but in only one place was the actual contact observed. On the northwest slope of North Mountain a small ridge projects westward beyond the rest of the monzonite. A contact of the Winnipesaukee gneiss (?) and the coarse-grained monzonite exposed on the south side of this ridge strikes N. 40°W. and dips 80°S.

This outward dip and that in the diorite are similar to the outward dip of the ring-dikes on the Island of Mull (Clough, Bailey, and Wright, 1924) and comparable to those of the New Hampshire ring-dike areas (Chapman, 1935; Kingsley, 1931; Modell, 1936), which have essentially vertical contacts.

Middle Mountain complex.—Middle Mountain forms a gently curving arc, concave to the northwest, which trends northeast, approximately parallel to the outer, southeast margin of the complex (Pl. 1). It is made up of coarse-grained and fine-grained monzonite in more or less distinct bodies, although distinct contacts between them are rarely exposed.

A core of coarse-grained monzonite separates the two larger bodies of fine-grained monzonite. Another body of coarse-grained monzonite forms the northeastern end of the mountain and extends across the road in the vicinity of the parking area. In a road cut at the entrance to the parking area a sharp contact between the coarse-grained monzonite and the small body of fine-grained monzonite indicates that here the coarse monzonite is younger.

In spite of textural variations and possible age differences the Middle Mountain complex is believed to constitute a single structural unit. As such it is consistent with the structure of the whole plutonic area, forming an arcuate mass concave to the northwest.

The small mass of porphyritic gabbro and augite diorite on the southeast side of Middle Mountain is believed to be an inclusion of the marginal phase of the earlier diorite body.

Dikes.—Any structural significance attributed to the dikes would imply that the dikes actually observed represent a majority of those present. Conditions of exposure preclude the possibility of finding dikes over most of the area, and the dikes mapped are those in favorably exposed areas. The writers hesitate to attach any significance to what they consider an inadequate sample.

The following general observations may be made regarding the dikes. The pegmatites, with one exception, are in the gneiss and are granitic pegmatites not associated with the intrusive complex. The exception is a basic pegmatite in the diorite along the road entering the area from the west. The aplite dikes observed are restricted to the area of the intrusive complex, whereas the monzonite and lamprophyre dikes occur in both the gneiss and the intrusive complex. A few dikes, the identification of which is uncertain, are shown without type symbol.

Joints.—Joints, like the dikes, may be observed adequately only in favorable areas, and the data are therefore meager. Two systems, one striking N. 35°E., the other N. 35°W., and both having steep (65°–90°) dips, seem to be dominant. Isolated observations show joints of random orientation, and essentially horizontal sheeting is present locally.

The joints are rather widely spaced and prominent enough to permit the sapping of many large blocks from North Mountain by Pleistocene continental glaciers. These boulders were deposited in the valley northwest of Round Pond and comprise the boulder field which extends from "Balanced Rock" and "Ballard Rock" upstream to the locally famous "Pawtuckaway Boulders." These boulders range up to tens of feet in greatest dimensions.

AGE RELATIONS

Before considering the mode of intrusion of the Pawtuckaway complex it is necessary to consider briefly the age relations of its component parts.

The oldest phases of the complex are the gabbros and diorites, inclusions of which are found in the monzonites. Although it is impossible to determine the relative ages of the outer monzonite arc and the Middle Mountain monzonite arc, they are both younger than the more mafic phases and their similarity in composition to each other suggests that they are essentially of the same age.

The first intrusion may well have been a normal diorite magma in which a marginal gabbro and augite diorite phase developed. This was followed, perhaps closely, by the intrusion of the monzonite bodies, one between the diorite and the older gneiss and the other almost in the center of the diorite mass. These major units were followed by the injection of a variety of dike rocks.

MODE OF INTRUSION

GENERAL STATEMENT

The method or methods by which magma bodies reach their observed levels in the earth's crust ranks with the physical chemistry of magma genesis as one of the leading problems in igneous geology. At present the emplacement of most igneous bodies is explained by: (1) forceful injection, (2) piecemeal stoping, (3) cauldron subsidence, or a combination of these. Although there is a scarcity of exposures along contacts the authors feel that available evidence indicates the essential mechanics of intrusion. It will be desirable to consider the evidence for and against each of the methods mentioned above, with respect both to the initial diorite magma and to the later monzonite bodies.

DIORITE BODY

Forceful injection.—The elliptical outline of the diorite body and the presence of strong foliation in part of it suggests forceful injection. Balk (1937, p. 92–94) has pointed out that many gabbroic stocks are funnel-shaped and that foliation, if present, will be nearly parallel to the inward-dipping walls. The foliation of the diorite strikes essentially parallel to the outline of the body and dips steeply inward as it should if the body is funnel-shaped. However, Chapman and Chapman (1940, p. 207–208) in their discussion of the intrusion of the very similar gabbro-diorite stock of the Ascutney Mountain complex point out what seems to be a fatal objection to the forceful injection of these bodies. The essential difficulty they find is that of raising the conical block of country rock to a sufficient height to clear the present erosion surface. If a conical fracture did develop above the diorite magma body, the subsequent invasion of the magma might lift the block in essentially the cone-sheet fashion. However, with free passage to the surface it would be difficult or impossible for the magma to lift the block far enough to permit the formation of a stocklike body and only a cone-sheet or funnel-shaped dike could result.

If the foliation in the diorite is a result of flow in a funnel-shaped fracture, it should

be more uniformly developed throughout the body. The localization of the foliation is better explained by another mechanism.

The one exposure of the diorite-gneiss contact shows that at least locally the walls of the intrusive dip outward rather than inward. The trace of the contact across topographic irregularities suggests strongly that the walls of the body have a steep outward dip throughout.

Although evidence against the diorite body being a forceful, funnel-shaped injection seems conclusive, there remains the possibility that it is a stock which forced aside the older rock to reach its present level. Field evidence, however, is strongly against this idea. The diorite body is elliptical in plan, and the greater dimension is along a general east-west direction almost normal to the trend of the schistosity in the older gneiss. Certainly, if the magma were injected into a yielding, schistose rock, the body should be elongated parallel to the regional structure. Furthermore, the foliation of the older gneiss should tend to wrap around the intrusion if it were forcefully injected, and this is not true.

Thus there seems to be adequate evidence that the diorite body is neither a funnel-shaped injection nor a forcefully injected stock.

Piecemeal stoping.—That many magma chambers are enlarged to some unknown extent by the stoping of small blocks from the sides and roof cannot be doubted. If inclusions of the older rock are abundant, stoping is beyond dispute; and if inclusions are rare or absent, as is the case here, it does not preclude stoping. The absence of inclusions of gneiss in the diorite make it impossible to evaluate the importance, if any, of stoping in determining the size and shape of the chamber. Without evidence against stoping as a major factor and with only circumstantial evidence in favor of cauldron subsidence, the authors nevertheless favor the latter as the essential mechanism for the emplacement of the diorite body.

Cauldron subsidence.—For reasons discussed below, with respect to the monzonite bodies, the authors favor cauldron subsidence as the method of intrusion of the diorite body.

MONZONITE BODIES

Forceful injection.—The general position of the monzonite bodies is obviously related to the form and position of the earlier diorite body, and therefore the mechanism of emplacement should be related. The arguments against forceful injection of the diorite body apply equally to the monzonites and need not be repeated.

Piecemeal stoping.—Although inclusions of the older gabbro and diorite occur in the monzonites, the authors assign an incidental role to stoping in the emplacement of the latter.

Cauldron subsidence.—The mechanism favored by the authors to account for the emplacement of the Pawtuckaway complex is that of cauldron subsidence or ring-fracture stoping. The following considerations individually support and collectively justify this conclusion:

(1) The lack of compelling evidence in favor of piecemeal stoping and the presence of evidence against forceful injection make necessary the adoption of some other mode of intrusion.

(2) The circular plan of the complex and especially the arcuate shape of the monzonite bodies is similar to the plans of ring-dike areas elsewhere in New England and in Scotland (Fig. 1). This in itself suggests that igneous complexes with such similar structural patterns should have a common origin.

(3) Petrographically the rocks of the Pawtuckaway complex belong to the White Mountain magma series, occurrences of which are uniformly circular or arcuate in plan, and geologists who have studied the other areas have all favored cauldron subsidence as the mode of intrusion. The present authors cannot explain this uniformity of plan, but the lack of explanation in no way detracts from the facts or the significance which we attach to them.

(4) Data indicate that the walls of the complex dip outward at rather steep angles consistent with the cauldron subsidence theory. This is true even of the diorite body although the internal structure of the diorite dips inward.

(5) The favored mechanism is the only one which allows a logical and not too involved interpretation of the intrusive history of the complex.

CONTRASTS WITH OTHER AREAS

In a number of respects the Pawtuckaway complex differs from other areas of ring-dike intrusion in New England:

(1) The dimensions are smaller than those of most of the bodies belonging to the White Mountain magma series.

(2) Volcanics are the earliest igneous rocks in most of the areas but are absent here.

(3) There are slight petrographic differences which were indicated above.

(4) A smaller variety of rock types is represented here; conspicuously absent are the quartz-bearing members of the series.

SEQUENCE OF EVENTS

The sequence of events in the intrusion of the Pawtuckaway complex is believed to be:

(1) The subsidence into a magma chamber of an essentially circular block bounded by fractures which dip steeply outward. The magma welled up around and over this block producing the diorite body with local gabbroic facies.

(2) Further subsidence of the block but with greatest reopening of the ring fracture on the south, east, and north sides, allowing the later magma to rise more freely in these areas. This would account for the outer monzonite arc. Although the relative ages of the outer and inner monzonite arcs are unknown, the arcuate character of the inner arc indicates that it resulted from the same mechanism as the outer arc, and the petrographic similarity of the two favors near contemporaneity.

The injection of the monzonite bodies along with the subsidence of the central area, including the diorite, is believed to be the cause of the foliation in the diorite. The coarse texture of the diorite suggests that the body had a long magmatic life, and it is believed that the diorite was still plastic enough to yield by flow when the marginal and central monzonites were intruded. This would not only afford an adequate stress to produce the foliation but would also explain the localization of the foliation to the portion of the diorite mass between the two monzonite arcs.

It may be questionable whether a magma may be liquid enough to yield by flow and at the same time rigid enough to fracture and form inclusions. The authors see no difficulty in having local, marginal areas readily fracture and thus produce inclusions, while the whole mass will flow and produce foliation.

RELATION OF INTRUSION TO DIFFERENTIATION

The magmatic differentiation of the White Mountain magma series has been discussed in detail by Chapman and Williams (1935). Their conclusions, based on studies of the then-known areas of these rocks, are equally valid for the Pawtuckaway complex. Although certain differences in the minerals have been noted, these differences are minor in relation to petrogenic problems.

The rocks of the Pawtuckaway complex become progressively less mafic from the earlier gabbroic intrusions to the latest monzonite and syenite. The dominant mafic mineral of the earlier rocks is augite, whereas in the later diorites and monzonites hornblende and biotite dominate. The prevailing plagioclase ranges from labradorite in the gabbroic rocks through andesine in the diorites to oligoclase in the monzonites. The accessory minerals, apatite and magnetite, are present in all the rocks and vary only slightly in amount.

The earliest intrusion was gabbroic, and the authors believe that this magma differentiated in place to produce the diorite body, although the latter may represent a later intrusion from a differentiating parent magma at depth. The monzonite bodies were intruded from a parent magma which had differentiated at depth. After intrusion the monzonite magma became syenitic locally but in general crystallized without appreciable modification.

CONCLUSIONS

(1) The Pawtuckaway complex belongs to the White Mountain magma series.

(2) The complex contains a variety of rocks ranging from gabbro to syenite but is composed mainly of diorite and monzonite.

(3) The structural character of the complex is best accounted for by the cauldron subsidence mechanism of emplacement.

(4) The rocks become progressively less mafic from oldest to youngest, which with other petrographic characteristics suggests a rather simple magmatic history.

REFERENCES CITED

Balk, Robert (1938) *Structural behavior of igneous rocks*, Geol. Soc. Am., Mem. 5, 177p.

Barnes, V. E. (1930) *Changes in hornblende at about 800°C*, Am. Mineral., vol. 15, p. 393.

Billings, M. P. (1928) *The petrology of the North Conway quadrangle in the White Mountains of New Hampshire*, Am. Acad. Arts Sci., Pr., vol. 63, p. 67–137.

———, (1937) *Regional metamorphism of the Littleton-Moosilauke area, New Hampshire*, Geol. Soc. Am., Bull., vol. 48, p. 463–566.

Chapman, R. W. (1935) *Percy ring-dike complex*, Am. Jour. Sci., 5th ser., vol. 30, p. 401–431.

———, and Chapman, C. A. (1940) *Cauldron subsidence at Ascutney Mountain, Vermont*, Geol. Soc. Am., Bull., vol. 51, p. 191–212.

———, and Williams, C. R. (1935) *Evolution of the White Mountain magma series*, Am. Mineral., vol. 20, p. 502–530.

Clough, C. T., Bailey, E. B., and Wright, W. B. (1924) *The Tertiary and post-Tertiary geology of Mull, Loch Ame, and Oban*, Geol. Survey Scotland, Mem.

———, Maufe, H. B., and Bailey, E. B. (1909) *The cauldron subsidence of Glen Coe*, Geol. Soc., London, Quart. Jour., vol. 65, p. 611–678.

Daly, R. A. (1933) *Igneous rocks and the depths of the earth*, McGraw-Hill, New York, 508 p.

Hitchcock, C. H. (1877; 1878) *The geology of New Hampshire*, vols. 2, 3, State of New Hampshire, Concord.

Jackson, C. T. (1844) *The geology of New Hampshire*, State of New Hampshire, Concord.

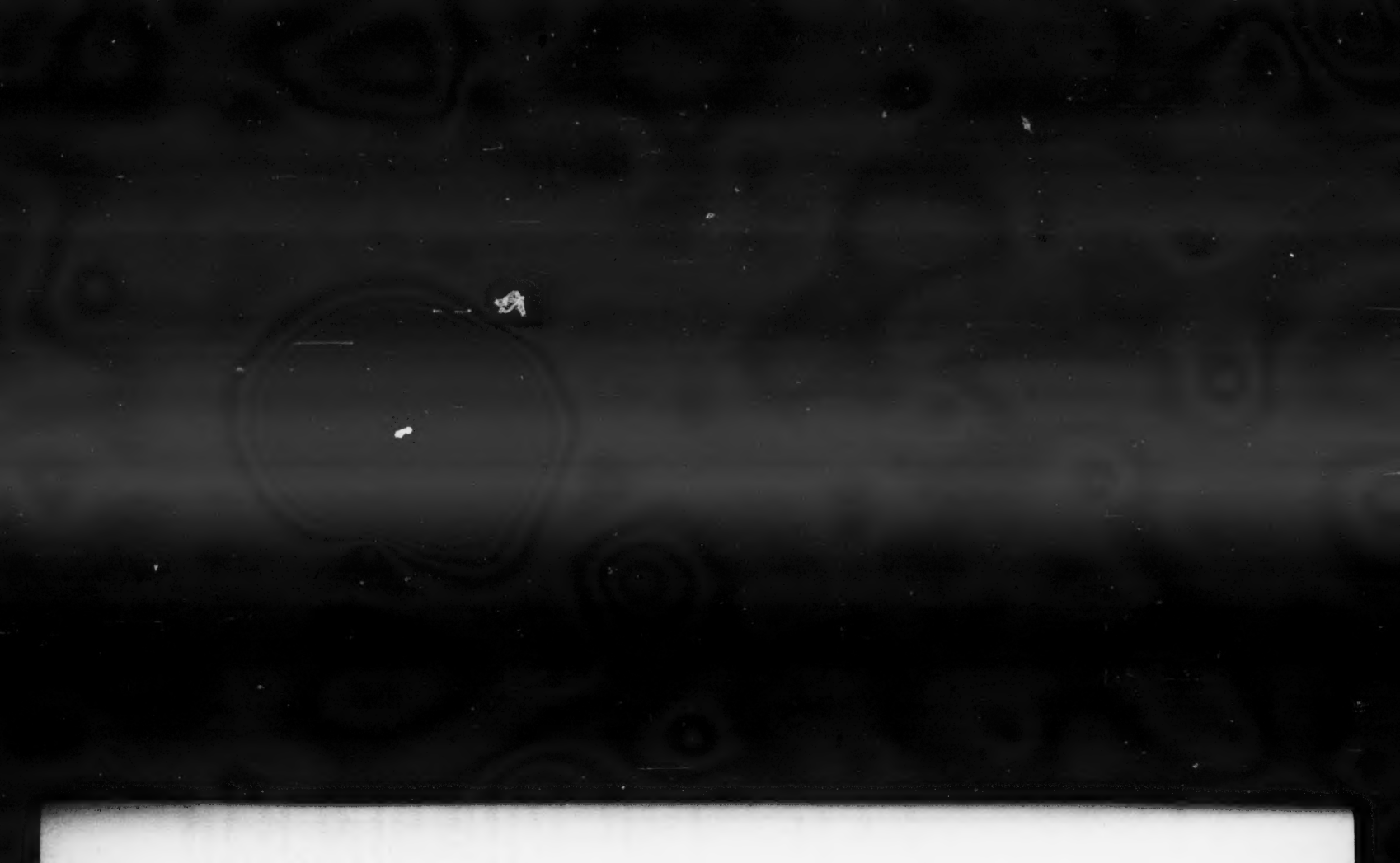
Kingsley, Louise (1931) *Cauldron-subsidence of the Ossipee Mountains*, Am. Jour. Sci., 5th ser., vol. 22, p. 139–168.

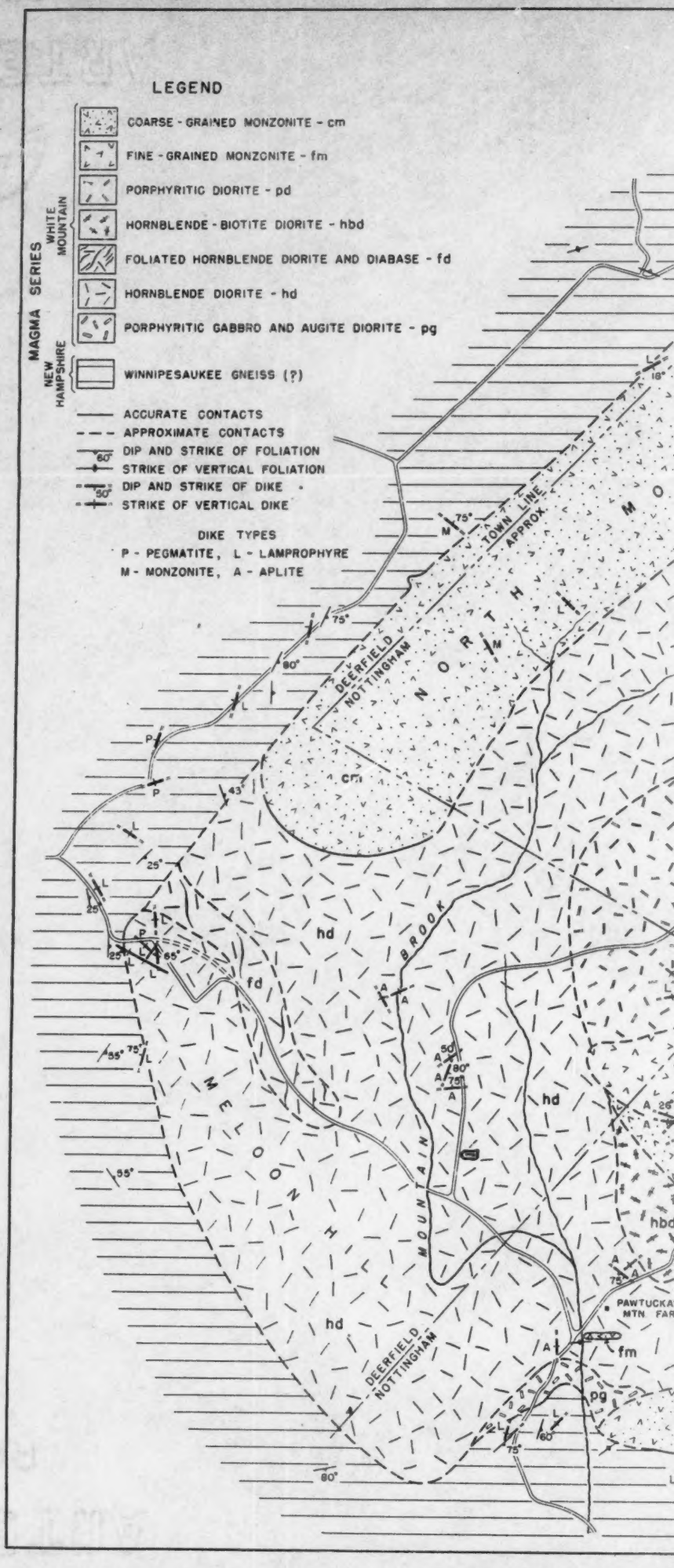
Modell, David (1936) *Ring-dike complex of the Belknap Mountains, New Hampshire*, Geol. Soc. Am., Bull., vol. 47, p. 1885–1932.

Smith, Althea P. (1941) *Olivine and pyroxene of Mt. Tripyramid, New Hampshire* (Abstract), Am. Mineral., vol. 26, p. 202–203.

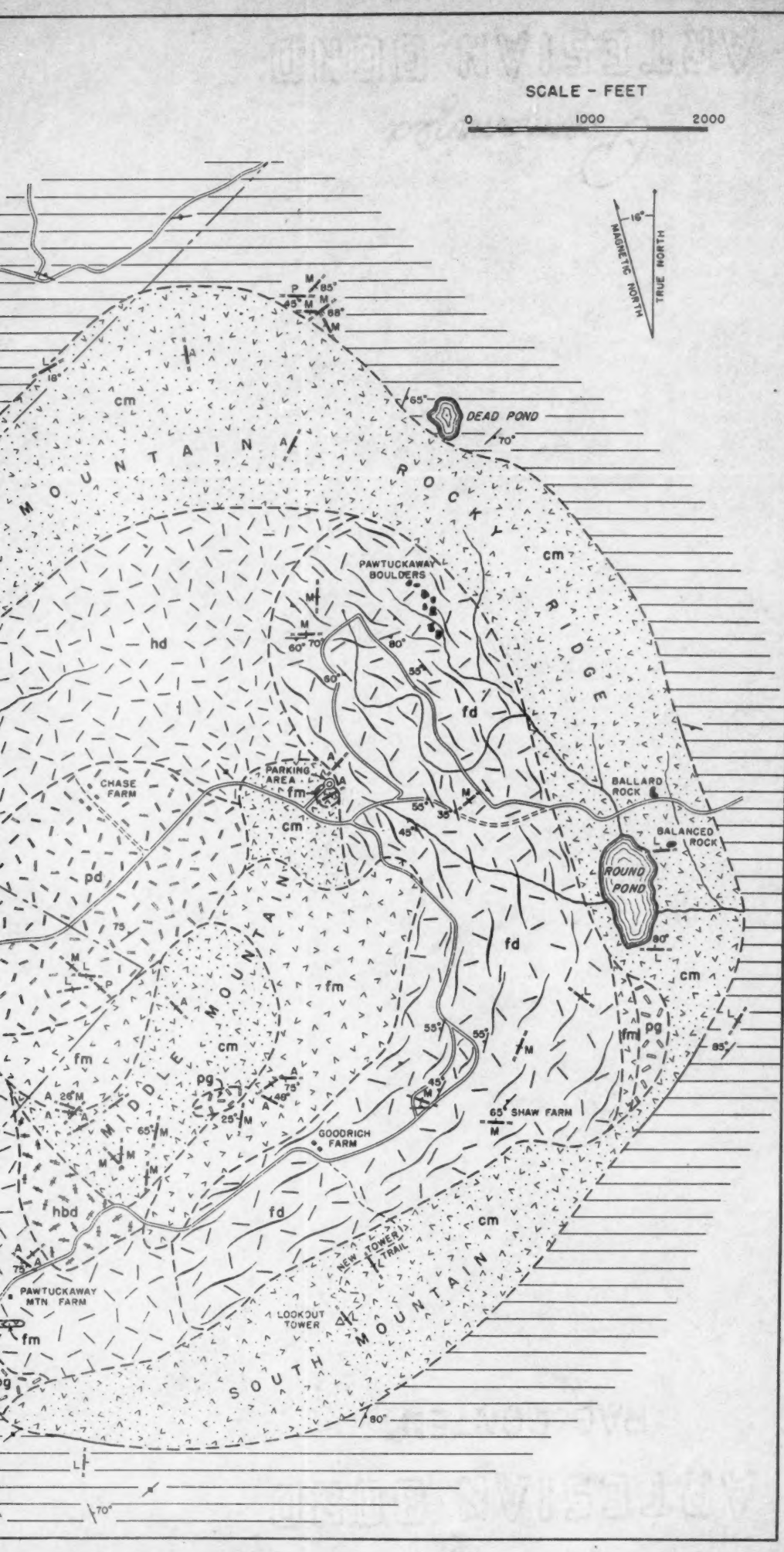
Smith, Edward S. C. (1922) *Geology of the Pawtuckaway Mountains, New Hampshire* (Abstract), Geol. Soc. Am., Bull., vol. 33, p. 128.

Williams, C. R., and Billings, M. P. (1938) *Petrology and structure of the Franconia Quadrangle, New Hampshire*, Geol. Soc. Am., Bull., vol. 49, p. 1011–1044.





GEOLOGIC MAP OF P



MAP OF PAWTUCKAWAY MOUNTAINS